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THE ORIGIN OF THE INCLUSIONS IN DIKES

SIDNEY POWERS
Geological Museum, Cambridge, Massachusetts

PART I

PART I

INTRODUCTION

Several dikes containing numerous inclusions have recently been seen by the writer, and the origin of these inclusions appears to be of sufficient interest to warrant a brief notice of these and other cases found in the literature.[1] The dikes have received various descriptive names, but none of them is sufficiently comprehensive to include all the examples.

Inclusions are not infrequent in all types of both extrusive and intrusive rocks. Those in extrusive rocks have been specially treated by Lacroix. Intrusive rocks are so extensive and of such varying forms that only dikes and a few related intrusions will be considered here. The xenoliths in the large intrusive bodies have formed the basis for the stoping hypothesis, but only part of the inclusions in dikes are of this origin.

Dikes acquire inclusions by shattering blocks off the walls of a fissure during their ascent through it. These fragments may remain near the place from which they came, or they may move up or down. In most cases the fragments rise, whether of greater specific gravity than the molten dike-rock or not, because they are forced upward by the magma. In the cases where the fragments sink, either they are heavier than the magma or they are carried down.

Some dikes have invaded conglomerates, from which they have dissolved the cement and included the bowlders. In other cases the fissures through which the magma came may have been open to the surface, so that stream gravels fell in and were caught in the ascending magma. In rare instances a dike may ascend through a fault breccia and thus acquire its inclusions. These special cases are not separated from the other examples.

To classify dikes according to the direction of movement of the inclusions is difficult, because in almost every case some inclusions rise and some remain near the place of origin. A classification is attempted which will at least serve to emphasize the number of examples in which inclusions sink.

[1] The writer is indebted to Professor R. A. Daly for suggestions concerning this paper.

EXAMPLES

(*A*) *All inclusions near place of origin.*—

Cornwall, England: In the southern part of Cornwall, near Carrick Luz, there is a gabbro boss surrounded by a fringe of gabbro dikes extending outward through the serpentines of the region. All of the dikes contain numerous inclusions of the serpentine. These dikes also illustrate injection foliation.[1]

Da la Bèche reported one of the granite dikes of this region to contain numerous fragments of the slate which it cuts.[2]

Mexico: In the state of Guerrero, Mexico, a granite sill from 12 to 15 feet wide cuts the Juratrias shales, and contains a number of fragments of the slate a foot in diameter. The vertical sill runs parallel to the bedding of the shales, but has evidently broken off the shale inclusions from the walls near the place where they are frozen in the dike. The inclusions are now nearly weathered out of the granite, so that the effect of the heating is indeterminate, but the granite adjoining them shows no immediate contact effect.[3]

Cape Ann, Massachusetts: At Pigeon Cove, Cape Ann, is a labradorite porphyry dike of very coarse grain, 25 feet wide and traceable for 3 miles, everywhere cutting alkaline hornblende granite, but including fragments of diabase and quartzite. The inclusions are about 6 inches in diameter and have a subangular to rounded outline. The numerous diabase dikes now exposed in the region cut the porphyry dikes. The inclusions must have come from quartzites and diabase dikes cutting them, forming the roof of the granite batholith, and must have sunk or been carried down in the dike to their present level. Since that time, which was probably the Carboniferous, the quartzites from the roof of the batholith and the upper part of the batholith itself have been completely removed by erosion. Quartzites appear in small amounts elsewhere in Essex County. The distance through which the inclusions must have

[1] J. S. Flett, "The Geology of the Lizard and Meneage," *Mem. Geol. Surv. Great Britain*, 1912; also, *Proc. Geologists' Association*, XXIV (1913), 127.

[2] *Geological Report on Cornwall* (1839), p. 182.

[3] The writer is indebted to Mr. Y. S. Bonillas for this information.

descended can only be estimated as probably several thousand feet.[1]

(*B*) *Some inclusions have sunk.*—

Montreal, Canada: A number of breccia dikes have been described from the vicinity of Montreal and the Monteregian Hills by Robert Harvie.[2] The dikes will be mentioned in the order in which they are described by Harvie.

Near La Trappe in the Oka Mountains are two inclusion-bearing alnöite dikes, one of which is enlarged to a width of 180 feet, while the other is 25 feet wide and traceable for half a mile. The fragments are of Grenville limestone, Laurentian gneiss, Potsdam sandstone, and Beekmantown sandstone. The last two formations are stratigraphically higher than the exposure which is near the contact of the Grenville and Laurentian. Therefore some of the blocks rose, others sank.

Near Ste. Anne de Bellevue is a dike a foot wide cutting Trenton limestone and containing angular fragments of sandstone, hornstone, and limestone. The sandstone has a quartzitic rim, the hornstone was apparently originally a shale, and the limestone is changed to a crystalline marble.

On Westmount Mountain at Little's quarry is a large breccia camptonite dike about 4 feet wide. In one side of the dike an offshoot from it "has eroded or stoped out and filled a large cavity." The heat of the intrusive has baked the limestone, and a section from the unaltered limestone to the dike shows increasing baking and shattering until the blocks of limestone drop off into the dike-rock, where they form the bulk of the angular inclusions. Harvie says:

> In this process of shattering by which the action of stoping went forward every gradation is visible, from the solid unaltered limestone to the fragments finally frayed off and held in suspension in the breccia. The circulation of the materials of the intrusive has evidently been very vigorous, since fragments of other rocks, granites, etc., are found carried up to the top of the stope within a foot of the unshattered limestone. The magma has not exerted any pro-

[1] The writer is indebted to Professor C. H. Warren, of Boston, for the information concerning the inclusions in this dike.

[2] On the origin and relations of the Paleozoic breccia in the vicinity of Montreal, see *Trans. Roy. Soc. Canada*, 3d ser., III (1909), 249–78.

nounced solvent action on the inclusions, either the basic limestone or the more acid granites; neither has the heat greatly affected the limestone, which one would expect to find in a crystalline state.

There are a number of other breccia dikes in this immediate vicinity, 2 to 7 feet wide, of both a camptonite and a pyroxenite composition. They contain fragments of granite, gneiss, essexite, syenite, Potsdam sandstone, and Trenton limestone, the country rock. All these inclusions have either come from below or from the limestone near by.

It is evident that the inclusions in the cases above mentioned have been derived almost wholly by the dike-magma shattering the walls as it advanced. There must have been an active circulation in the dikes to keep them at a high temperature at the top while they made their way through the limestone and probably through the underlying rocks. This circulation carried up the blocks of the pre-Cambrian and carried down the blocks of the Potsdam and Beekmantown. These dikes appear, therefore, to be cases where the ascent was due, at least partly, to a blowpiping action on the rock overhead.

In the same memoir Harvie describes other dikes and intrusions of various forms, all of which consist of a breccia of similar inclusions in a basic matrix. One of the masses of breccia on St. Helen's Island,[1] near Montreal, cutting Utica shale, contains blocks of fossiliferous Oriskany limestone and Helderberg limestone, indicating the former presence of strata of these ages in the region. It also contains fragments of the Ordovician and older rocks.

The vertical movement of the fragments in these cases is interesting because it has been in both directions in the same dike or mass. This movement does not seem to be dependent on the form of the intrusion nor on the composition of the intruding dike-rock. The thickness of the formations from the top of the Laurentian to the bottom of the Utica is 2,500 feet (Harvie). The Utica is succeeded by Lorraine shales 2,000 feet thick. No higher stratigraphic horizon is exposed in the region, except in the inclusions on St. Helen's Island. Therefore these Devonian fragments must

[1] See also F. D. Adams, *Twelfth Internat. Geol. Cong., Guide Book 3*, p. 55, for a description of these breccias.

have sunk or been carried down over 2,000 feet, and the Laurentian fragments in the same breccia brought up over 2,500 feet. In the La Trappe dike the movement in each direction was probably not as great, but the distance from the Beekmantown down to the contact of the Grenville and Laurentian must have been several thousand feet.

Marblehead, Massachusetts: At Marblehead, Massachusetts, half-way between Peaches Point and Naugus Head, are a number of dikes of pulaskite cutting an igneous complex. These dikes range in width from a few inches to 50 feet. In many cases they appear to follow the course of earlier diabase dikes, all stages being present, from a network of pulaskite veins in a solid diabase dike to a shatter breccia of diabase in pulaskite and finally to a band of schistose streaks of diabase in the center of a broad dike of pulaskite, showing the former position of a diabase dike. In one case there are remnants of two such dikes in a wide pulaskite dike. The diabase fragments are lenticular schistose bands, sometimes several inches in width, in a coarse-grained pulaskite, which grows finer-grained in the small tongues which permeate the diabase inclusions. The dark constituents of the pulaskite appear to have been derived largely from the absorbed diabase. The width of the original dikes was from 1 to 3 feet.

Southern Sweden: At Brevik and at Karlshamn, southern Sweden, diabase dikes filled with rounded inclusions occur. They have been described by Hedström[1] and Eichstädt[2] on whose papers the following description is based. Two of these dikes are known, being mapped on the Eskjö and Karlshamn sheets of the Swedish Geological Survey.

The dike at Karlshamn is 3 miles long and in places several hundred feet wide. The inclusions are all on the west side of the dike and consist wholly of quartzite and quartzitic sandstone. Cataclastic structure is characteristic of all the inclusions, but this texture was caused by metamorphism before the quartzites were

[1] "The Pebble-Diabase of Brevik," *Eleventh Internat. Geol. Cong., Guide Book 18*, 1907, pp. 47–51.

[2] "Om quartsit-diabas-konglomeratet i Småland och Skåne," *Sveriges Geol. Unters.*, Ser. c, No. 74, 1885.

included in the dike. The contact of the pebbles and the diabase in the case of this dike is sometimes sharp and sometimes indistinct. In the pebbles, which have undergone the most marked absorption, it is difficult to distinguish even the center of the pebble. Sometimes ilmenite and fine feldspar crystals have formed in the pebbles with a rim of epidote and chlorite between the pebble and the diabase. The dike cuts Archean gneiss, and none of the younger pre-Cambrian sedimentary series is present in the region.

The "pebble-diabase" dike of Brevik (on the Eksjö sheet) has a strike of about N.–S. and has been traced for about 15 miles. It cuts Almesåkra pre-Cambrian sandstone, quartzite, and conglomerate. The width of the dike varies in different exposures and it often exceeds 300 feet. Inclusions appear only in some of the outcrops, and there they are, as a rule, confined to smaller zones, often 10 to 15 feet in width, which are elongated parallel to the sides of the dike. These zones may be situated either on the sides of the dike, and usually on the eastern side, or near the middle. They vary in width from 4 to 50 feet, changing with the width of the dike. In smaller offshoots from the dike, about 15 feet in width, fragments sometimes occur evenly distributed over the entire width. Elsewhere there is a sharp boundary between the parts free from and full of inclusions.

The inclusions consist largely of quartzite and schist derived from the Almesåkra pre-Cambrian complex, with some granites, leptites, gneisses, and other pre-Cambrian rocks. The size of the fragments varies from a few inches to 30 feet. The shape of the quartzite inclusions is rounded, of the granite, subangular, while the inclosures of schist form thin bands of considerable length, which are surrounded by sheets of diabase. The pebbles are so numerous as almost to touch. They occupy about half the volume of the dike, and at times even more. The inclusions weather out easily.

In the pebble-bearing parts of the diabase the ophitic diabasic structure is not developed, and the presence of the numerous inclusions has caused a segregation of the light and dark minerals into separate spots. There are also grains of quartz and feldspar scattered through the diabase as if derived from the resorption of the edges of some of the inclusions or from the cement of the original

conglomerate from which the pebbles were derived (Hedström's theory). The inclusions of quartzite are generally surrounded by zones in which iron ore, biotite, and chlorite are developed. The contact of the diabase with the quartzite pebbles is, however, usually sharp, while the contact with the sandstone and granite is more or less indistinct, the magma having formed fused contacts and even having penetrated a few of the inclusions. In places a flow structure is developed in the diabase around the fragments.

Eichstädt proposes the theory that the inclusions in both dikes were derived from loose pebbles on the surface of the ground, probably accumulated in valleys, which fell into fissures formed in advance of the diabasic intrusions and were caught in the magma when it ascended in the fissures. In support of this theory he points out that if the blocks of quartzite were originally angular and the edges resorbed where they are now, the silica content of the diabase near the zones of inclusions would be much greater than elsewhere in the dike. This is not the case, although he finds free quartz and micropegmatite present in the diabase of both dikes. He noted the existence of pre-Cambrian conglomerate in the vicinity, but considers that the pebbles were derived from a loosely consolidated conglomerate (as this Almesåkra conglomerate may have been in the pre-Cambrian time when the dikes were formed) and not from a massive conglomerate.

Hedström proposes a somewhat similar theory: that larger and smaller pieces of the Almesåkra conglomerate have been imbedded in the diabase magma. He considers that the conglomerate was at least somewhat consolidated and accounts for part of the matrix by the numerous grains of quartz and feldspar scattered in the diabase. In support of this view he finds inclusions to which portions of the original matrix are still attached. The hard quartzite pebbles in the Almesåkra conglomerate are greatly cracked and show the same cataclastic structure as do those of the inclusions which have often split up into several pieces which have been more or less widely displaced from each other in the magma. The cement of the original conglomerate was gritty, loose, and brittle, and therefore its cohesion was easily destroyed at the intrusion of the diabase. "In the appearance and character of the pebbles, as

also in the proportions of the rock types represented among them, is a perfect agreement between the Almesåkra conglomerate and the pebble-diabase." It should be noted, however, that the Almesåkra series is now present only near Brevik. A former extension of the series for 90 miles south, to Karlshamn, may have existed.

Cripple Creek, Colorado: At Cripple Creek a number of phonolite and rhyolite dikes include or are capped by loose conglomerate and volcanic breccia. These phenomena have been described by G. H. Stone[1] who has confounded the origin of the conglomerate and breccias with that of the dikes. As pointed out by Ransome and Lindgren,[2] the former are stream gravels and volcanic breccias, some of which have been invaded by dikes. It appears probable that the inclusions in the dikes themselves have fallen in from the overlying loosely consolidated beds.

At Grizzly Peak, Colorado, similar granitic breccias have been reported by G. H. Stone[3] and it is probable that here also stream gravels or volcanic breccias have been invaded by dikes.

The manner in which an igneous rock invades a conglomerate is well illustrated in a satellitic stock of the Bayonne batholith in the Selkirk Mountains, British Columbia, as described by R. A. Daly.[4] The granitic magma has eaten its way into the conglomerate,

> dissolving out the cement in large amount, and has thus not only thoroughly impregnated the conglomerate with the granitic material, but has quite separated many of the larger quartzitic pebbles, which, still rounded, are now completely inclosed in granite. The cement was evidently more soluble in the magma than were the quartzite pebbles—a conclusion to be expected in view of the fact that the heterogeneous cement has a lower fusion-point, and in relation to the acid granite, a lower solution-point of temperature, than the more highly siliceous quartzite. The partial absorption of the conglomerate must have taken place when the magma was (because cooled down) sufficiently viscous to allow of the suspension of the blocks and pebbles. At an earlier period, when the cooling was less advanced, the quartzite pebbles

[1] The Granitic breccias of the Cripple Creek Region," *Amer. Jour. Sci.*, Ser. 4, V (1898), 21–32.

[2] *U.S. Geol. Surv., Professional Paper 54.*

[3] *Amer. Jour. Sci.*, Ser. 4, VII (1899), 184–86.

[4] "The Geology of the N. A. Cordillera at the 49th Parallel," *Can. Geol. Surv., Memoir 38* (1914), p. 300.

themselves, like the main quartzitic and schistose formations, could have been dissolved.

Pequawket Mountain, New Hampshire: Pequawket Mountain, the Eastern Kearsarge of New Hampshire, and Moat Mountain, near by, are composed of masses of quartz porphyry, of stock-like form, which contain so many fragments of metamorphic rocks that they have been described as breccias.[1] The Pequawket mass is about 1,200 feet long and 450 feet wide. It lies at the contact of the older Albany granite and slate. The angular inclusions are very numerous, consisting largely of slates, sandstones, and phyllites, derived from the adjacent terranes. The inclusions do not show any alteration at their contact with the porphyry. The porphyry matrix is vitrophyric in the Pequawket mass, granophyric in the Moat mass. As the inclusions do not show as much metamorphism as the surrounding slates, and as the quartz porphyry is not squeezed, Daly concludes that the inclusions have probably come from above the present exposure, where the metamorphism was not so great.

Lages, Brazil: Near Lages, Brazil, an inclusion-bearing dike was found by Woodworth.[2]

The dike is of trap and was evidently the feeder of one or more of the Triassic trap sheets which are exposed in the Lages area. At the outcrop investigated, the width of the dike is 75 feet. Here the inclusions of foreign rocks constitute one-half of the volume of the dike and comprise red and black shale, coarse-grained basalt, fine-grained basic rock, and amygdaloidal basalt. The fragments of sedimentary rock were apparently disrupted from the walls of the fissure which appears to have been a fault-line. Whether they came from above or below the exposure could not be determined. The amygdaloidal basalt fragments must have come from the overlying Triassic flows according to the determination of the geological structure by Woodworth and others. Furthermore, it is certain that the dike was not formed until after other lavas had been extravasated and cooled to yield the inclusions which sank in the fluid dike-magma.

[1] R. A. Daly, *Science*, N.S., III (1896), 752.

[2] J. B. Woodworth, "Geological Expedition to Brazil and Chile, 1908–9," *Bull. Mus. Camp. Zoöl., Harvard,* LXI, No. 1, p. 95.

THE MIDDLE AND UPPER DEVONIAN OF THE ROMNEY, WEST VIRGINIA, REGION[1]

CHARLES S. PROSSER
Ohio State University

CONTENTS

INTRODUCTION

The name Romney shales, from "Romney in West Virginia," was published by N. H. Darton in 1892 "for the basal series of dark shales [in the Devonian]."[2] The formation was briefly described as follows:

> The basal members [of the Devonian] are fissile shales, in greater part black or dark brown in color, containing occasional thin beds of sandstone and limestone. Their average thickness is about 600 feet. . . . . The Devonian formations are not fossiliferous at many horizons in the region west of Staunton. In the Romney shales the following species are Corniferous [misprint for conspicuous]: *Discina lodensis*, *D. minuta*, *Orthis leucosia*, *Strophodonta demissa*, *Cyrtina hamiltonensis*, *Spirifera mucronatus*, *S. granulifera*, and *Leiorhynchus limitaris*. This is a Hamilton group fauna, but the stratigraphic range of Hamilton group equivalents in the Romney shales is not apparent, and Hamilton deposits probably extend some distance above.[3]

[1] Published by permission of Dr. William Bullock Clark, state geologist of Maryland.

[2] *American Geologist*, X (July, 1892), 17, and the name first appears in the table of formations on p. 13.

[3] *Ibid.*, pp. 17, 18.

The Staunton folio by Darton, which appeared in 1894, was the first one published by the United States Geological Survey for the Appalachian part of the Virginias. In this folio Darton gave the following description of the Romney shales:

> The rocks consist of dark shales, black and fissile below, but somewhat lighter and more compact above. Some of the basal beds are carbonaceous to a moderate degree, and they have been worked at several points with the mistaken idea that they might prove to be coal-bearing. The formation includes occasional calcareous streaks not far from its base, and the upper members contain alternations of thin, pale-brown or dark-buff sandy beds, which constitute beds of passage into the next succeeding formation. The vertical range and stratigraphic position of these passage-beds appear to be somewhat variable, so that there is no definite line of demarcation between the two formations.[1]

In both of these publications the Romney shales were limited below by the Monterey sandstone, which later has been shown to be the southern continuation of the Oriskany sandstone of New York,[2] as was stated many years earlier by Hall.[3] The upper limit of the Romney shales was the base of the Jennings formation, which has been shown to correspond with the Genesee shale, Portage formation (including the Sherburne, Ithaca, and Enfield members), and Chemung formation.[4]

While Chief of Division of Appalachian Geology of the Maryland Geological Survey the writer studied with some care the outcrops in the vicinity of Romney, Hampshire County, in northeastern West Virginia. Particular attention was given to the exposures of the Romney shales, while the subjacent and superjacent formations were also examined, and because Romney is the typical locality for this formation it is believed that a somewhat fuller account than has yet been published is desirable.

[1] *Geologic Atlas of the United States*, Folio 14, p. 2, col. 4.

[2] Schuchert, *Bulletin Geological Society of America*, XI (1900), 271, 312–15; Prosser, *Journal of Geology*, IX (1901), 416; Schuchert, *Proceedings United States National Museum*, XXVI (1903), 420; and Rowe, Schuchert, and Swartz, *Maryland Geological Survey, Lower Devonian*, text (1913), p. 90.

[3] Geological Survey of New York, *Palaeontology*, III (1859), 40, 401.

[4] Prosser, *Journal of Geology*, IX (1901), 419–21; Prosser, *Maryland Geological Survey, Middle and Upper Devonian*, text (1913), pp. 345–49; Swartz, *ibid.*, pp. 423–34.

DESCRIPTION OF WEST VIRGINIA SECTIONS

Cliffs near South Branch of the Potomac River.—The nearest outcrops of any considerable extent are the cliffs of argillaceous blue pencil shales by the side of the road southwest of Romney and near the bridge over the South Branch of the Potomac River. Most of these shales contain small iron-like concretions and on weathering are generally stained a brownish or iron color. There is a decided tendency in most of them to split into narrow, long, pencil-shaped pieces. The cliff in places is from 20 to 25 feet in height; the cut continues for some distance and affords a fine exposure of this part of the Romney shales. The dip is not great at this locality, and soon after crossing the river to the west the Oriskany sandstone is reached.

Fossils are rare, but a few occur in some of the layers. *Liorhynchus limitare* (Vanuxem), *Pterochaenia fragilis* (Hall), *Productella Ambocoelia,* a Crinoid stem, and perhaps a few other species occur. *Pterochaenia fragilis* (Hall) is known to occur in New York from the Marcellus shale to the Chemung formation inclusive, while *Liorhynchus limitare* (Vanuxem) is confined to the Marcellus shales and regarded as one of its most characteristic fossils. These bluish to blackish shales, containing the characteristic species—*Liorhynchus limitare* (Vanuxem)—of the Marcellus shale of New York, occur in at least the lower half of the Romney formation and are correlated with the Marcellus shale of New York.

Outcrops from Romney to Hanging Rock.—The valley road between Romney and Hanging Rock, which is down the river to the north of Romney, affords frequent opportunities to study the Romney formation, and the exposures may be considered as typical for its upper part.

Ledges of bluish-gray to grayish shale occur along the road by the side of Big Run, about one-half mile north of Romney, some of which has a decided concretionary tendency, forming large irregular layers. Fossils occur abundantly in certain of these layers, especially species of *Chonetes* and *Camarotoechia,* the latter on account of their abundance forming layers that are

partly calcareous. The following species were obtained at this locality:

Chonetes mucronatus Hall
Chonetes scitulus Hall
[The specimens are larger than the figures of this species, but in this respect do not differ from New York specimens in the Hamilton which are so referred.]
Chonetes lepidus Hall
Productella sp.
Camarotoechia congregata (?) (Conrad)
Camarotoechia sappho (?) Hall
Tropidoleptus carinatus (Conrad)
Cyrtina hamiltonensis Hall
Spirifer mucronatus (Conrad)
Spirifer granulosus (Conrad)
Ambocoelia umbonata (Conrad)
Vitulina pustulosa Hall
Grammysia sp.
Nucula corbuliformis (?) Hall
[External and internal impression.]
Nucula varicosa (?) Hall
Palaeoneilo constricta (Conrad)
Modiomorpha concentrica (Conrad)
Pleurotomaria (*Bembexia*) *sulcomarginata* Conrad
Macrochilus hamiltoniae Hall
Orthoceras sp.
Crinoid stems.

Dr. Kindle also stated that Hamilton fossils "may be collected near the railroad station at Romney."[1]

At the next rock-cut on the highway, about one mile north of Romney, the shales vary in color from bluish gray to gray and are rather more argillaceous than those of the preceding exposure. A number of small iron-like concretions occur in them, but fossils are not nearly so abundant as at the following locality and in some of these shales they are very rare. Toward the southern end of the bank they are more abundant and species of *Chonetes*, *Bellerophon*, *Orthoceras*, and some other forms occur. The complete list follows:

Chonetes mucronatus Hall
Chonetes scitulus Hall
Ambocoelia umbonata (Conrad)

[1] *United States Geological Survey, Bulletin 508*, p. 41.

Nucula bellistriata (Conrad)
Palaeoneilo constricta (Conrad)
Bellerophon sp.
Bellerophon (*Patellostium*) *patulus* (?) Hall
Loxonema hamiltoniae (?) Hall
Orthoceras sp.
Orthoceras constrictum (?) Vanuxem
Spyroceras crotalum (?) (Hall).

By the side of the highway about $2\frac{1}{4}$ miles north of Romney are outcrops of blue, arenaceous shale, the dip of which is very heavy, being nearly 80° east. These shales split into medium-sized pieces which do not disintegrate very rapidly into soil. The color is distinctly bluish and there are calcareous bands composed largely of fossils, *Spirifer mucronatus* (Conrad) being the most abundant species. Other Brachiopods are common, as *Tropidoleptus carinatus* (Conrad), *Chonetes*, etc., while there are also some small Lamellibranchs and Gastropods, but the species of these last two classes are rare. It is essentially a Brachiopod fauna of the Hamilton formation. The complete list of the species obtained at this locality is given below:

Stropheodonta (*Leptostrophia*) *perplana* (Conrad)
Chonetes mucronatus Hall
Chonetes scitulus (?) Hall
Chonetes setiger (Hall)
Chonetes lepidus Hall
Chonetes vicinus (Castelnau)
Productella sp.
Tropidoleptus carinatus (Conrad)
Cyrtina hamiltonensis Hall
Spirifer mucronatus (Conrad)
Vitulina pustulosa Hall
Grammysia lirata (?) Hall
Nucula randalli (?) Hall
Pterinea flabellum (?) (Conrad)
Aviculopecten sp.
Pleurotomaria (*Bembexia*) *sulcomarginata* Conrad
Loxonema hamiltoniae Hall
Tentaculites bellulus Hall
Orthoceras subulatum (?) Hall
Dalmanites (*Cryphaeus*) *boothi* (Green)
Crinoid segments.

This exposure is very similar to hundreds of outcrops in the arenaceous Brachiopod zones of the Hamilton shales in New York, and any geologist who has studied that formation in New York would at once recognize the great similarity in lithologic appearance. This, together with the presence of a Hamilton fauna, apparently justifies the writer in correlating that part of the Romney formation shown in the highway cuts north of Romney with the Hamilton formation of the standard New York section.

Hanging Rock.—About four miles north of Romney is a high cliff known as Hanging Rock, where the South Branch of the Potomac River has cut a deep and narrow gorge from the east to the west through Mill Creek Mountain. The river makes a big loop, and at the highway bridge south of Springfield it turns and cuts back through the ridge to its eastern side. At Hanging Rock, on the northern side of the river, is a high cliff in which the anticlinal arch of the fold in the Oriskany sandstone is finely shown. A fair idea of the appearance of this cliff may be gained from the halftone (Fig. 1.) At the western end of the gorge, by the side of the highway just after crossing the railroad track, is an excellent exposure of fine black to drab shales containing small concretions and some fossils. In one layer at least are numerous specimens of *Phacops cristata* Hall associated with *Dalmanella, Chonetes, Ambocoelia* and the representatives of a few other genera. The rocks are slightly arenaceous and break into small, somewhat elongated pieces on weathering. The layer in which the fossils occur abundantly is somewhat arenaceous and not very much above the top of the Oriskany sandstone. The species collected in this shale are given in the following list:

Stropheodonta sp.
Chonetes cf. *lepidus* Hall
    [Two small imperfect internal impressions.]
Dalmanella lenticularis (Vanuxem)
Cyrtina hamiltonensis (?) Hall
    [Small imperfect specimen.]
Ambocoelia umbonata (Conrad)
Styliolina fissurella (Hall)
    [Abundant on some of the blocks.]
Phacops cristata Hall

In the above fauna, *Dalmanella lenticularis* (Vanuxem) is confined to the Onondaga formation; *Cyrtina hamiltonensis* Hall occurs in the Onondaga, Hamilton, and Portage; *Ambocoelia umbonata* (Conrad) from the Onondaga to the Chemung inclusive; *Styliolina fissurella* (Hall), southern Onondaga shale, Marcellus, Genesee, and Portage black shales; *Phacops cristata* Hall, Oriskany of Ontario, elsewhere the Onondaga. These species all

FIG. 1.—Hanging Rock, West Virginia, where the South Branch Potomac River has cut through Mill Creek Mountain. The rock cliff is Oriskany sandstone.

occur in the shales from Pennsylvania across Maryland and the Virginias which Dr. Kindle has correlated with the Onondaga of New York,[1] and since they came from the same stratigraphic horizon—the lower part of the Romney shale not far above the Oriskany sandstone—it appears that this part of the Romney ought to be correlated with the Onondaga. This correlation is in agreement with Dr. Kindle's statement that "the Onondaga shale

[1] *Journal of Geology*, XIX (1911), 97; *United States Geological Survey, Bulletin 508*, 1912; and *Maryland Geological Survey, Middle and Upper Devonian*, text (1913), pp. 48, 49, and 53–59.

member may be seen along the road through Mill Creek Mountain Gap, 1½ miles southwest of town [Romney]. Here the section shows an abrupt transition from the gray, highly calcareous coarse sandstone full of Oriskany fossils to drab shale approximately 100 feet thick, with some calcareous bands containing the Onondaga fauna. Fossils are not so readily found as at many other localities. Among those collected are *Anoplotheca acutiplicata*, *Dalmanella lenticularis*, and *Ambocoelia umbonata*."[1]

It is believed that the above account shows conclusively that the Romney shales in their typical locality near Romney, West Virginia, represent the southwestern continuation of the Onondaga limestone, Marcellus shale, and Hamilton formation of New York.

Outcrops south and north of Springfield.—North of Hanging Rock the valley road crosses a ridge of Oriskany sandstone and then for some distance south and north of Springfield the outcrops are mostly of thin black shales closely resembling lithologically the Marcellus shale member of the Romney formation. North of Springfield and perhaps three miles south of Green Spring in the cut of the Romney branch of the Baltimore and Ohio Railroad are black, argillaceous shales, and interstratified with the shales are several limestone layers, some of which are six inches in thickness. The limestone is very dark gray and blue to almost black in color, with a decided petroleum odor. It contains iron pyrites and is crossed by joints filled with calcite. There is a dip to the west and also to the south, and about 20 feet of rock is shown in the cut. Some of the layers of limestone contain a large number of small fossils, especially of the small species of *Ambocoelia*, which the writer has named *A. virginiana*.[2] Part of the limestone contains large numbers of this species, the best specimens of which came from this cut. When studied in the field these shales were referred to the Marcellus member; but it is possible that they occur lower in the Romney shale in that part of the formation which later has been correlated with the Onondaga.

Outcrops southeast of Romney.—To the east of Romney is the Jennings formation, succeeded by the Hampshire formation, the

[1] *United States Geological Survey, Bulletin 508*, p. 41.

[2] *Maryland Geological Survey, Middle and Upper Devonian*, text (1913), p. 202.

latter named from exposures in this county. There are only poor outcrops of the Jennings along the highway east of Romney, but farther east, in ascending the western slope of the northern part of South Branch Mountain, are excellent outcrops of red argillaceous shales belonging in the Hampshire formation. By the side of the highway, a short distance east of the first bridge over Big Run, are outcrops of sandstones and shales. The top and bottom layers consist of somewhat shattered, greenish to greenish-gray micaceous sandstone which breaks into blocks. Between the sandstones are thin, fissile, argillaceous shales, principally reddish-brown in color. The dip varies in different parts of the exposure from 11° to 14° S.E. No fossils were found, and the ledge is probably in the lower part of the Hampshire formation.

The rocks exposed by the side of the road east of this locality and up the valley of Big Run are mainly argillaceous shales, though there are some green ones, and interstratified with all of them are sandstones which are coarse-grained, greenish-gray in color, and massive. These outcrops occur along the highway below and in the vicinity of the locality known as the Peach Orchard, where, on top of the hill, the rocks are mainly red shale and the soil red in color, owing to their decomposition. These outcrops by the road and on the hill furnish a typical exposure of the Hampshire formation in northeastern West Virginia.

In Darton's account of the Devonian formations of central Virginia he says that "The Hampshire formation has yielded only a few plant remains which throw no light on the equivalency of the formation, but no doubt it comprised the representatives of the Catskill in their entirety or in greater part."[1] As stated by Darton, fossils are rare in this formation, as is the case in the corresponding one in Pennsylvania and New York; but the lithologic appearance and stratigraphic position agree, in general, with those of the Catskill formation, which has been shown to be a local one in New York, scarcely represented in the southwestern part of the state, while in the southeastern or Catskill Mountain region it has replaced all of the Chemung and the greater part of the Portage of western New York. It has been further shown that to the east of the

[1] *American Geologist*, X, 18.

Susquehannah River, in New York, the red rocks at first are interstratified with those which contain Chemung fossils, while farther east the red and greenish shales and sandstones replace all the rocks with the lithologic characters of the Chemung and its fauna disappears. Below the Chemung and in the midst of what corresponds to the Portage stage of western New York is another mass of red and greenish shales and sandstones called the Oneonta formation, which extends west to the Chenango Valley. In Delaware County the upper reds of the Oneonta and the lower of the Catskill unite and, in the Catskill Mountain region, extend downward into what is called the Sherburne sandstones, which represent the lower part of the Portage stage of western New York. In Pennsylvania, on following this mass of red rocks to the southwest it is found that they begin later, the change being gradual, and that the faunas of the Portage, including the Ithaca in Maryland and Chemung stages, reappear, as is the case in the southern part of western New York.[1]

The rocks exposed along the road following the highway toward Adams Mill, West Virginia, are largely red shales with some sandstones. Near the top of the hill are conglomerate layers, and across the upland the rocks are mostly red argillaceous shale, and all belong in the Hampshire formation.

In general it may be said that to the east of Romney the rocks belonging in the belt of the Jennings formation are mostly covered, while on the slope of the hill following the roads leading toward Frenchburg and Adams Mill are frequent outcrops of red shale interstratified with red sandstones belonging in the Hampshire formation. On top of these hills the red shales are most conspicuous, and decomposing rather readily into soil make good farming land.

Mill Creek southwest of Romney.—Mill Creek has cut a deep and narrow gorge through Mill Creek Mountain to the southwest of Romney which is followed by the highway from Romney to Moorefield Junction. In this gorge are high cliffs of the Oriskany sandstone which show excellently the massive nature of the formation.

[1] *Seventeenth Annual Report State Geologist* (New York), in which the writer has discussed this question.

The high cliff of Oriskany sandstone at the eastern end of the gorge is shown in Fig. 2. By the roadside not far west of the eastern end of Mill Creek Mountain Gap are layers of blue limestone interstratified with black argillaceous shales weathering to drab or brownish color. There are some fossils in the limestones, but they are infrequent. Along the road through this gap Dr. Kindle reported approximately 100 feet of drab shale with some calcareous bands succeeding the Oriskany sandstone, carrying an Onondaga

FIG. 2.—Eastern end of Mill Creek Mountain Gap, southwest of Romney, West Virginia. The steep cliff is composed of Oriskany sandstone.

fauna, and these basal shales of the Romney formation he correlated with the Onondaga of New York.[1]

Dr. Kindle further stated that: "The black shale representing the Marcellus is well exposed on the opposite side of the South Branch of Potomac River and appears to be barren. A little higher up the hill to the eastward on the west side of Romney the black fissile shale is succeeded by olive to gray slightly sandy shale with Hamilton fossils."[2]

[1] *United States Geological Survey, Bulletin 508*, p. 41.

[2] *Loc. cit.*

On the Mill Creek Road, $5\frac{3}{4}$ miles southwest of Romney or $1\frac{1}{4}$ miles east of Moorefield Junction, on the farm of Mr. Parker, is an outcrop of argillaceous shale. The dip is 35° N., 30° W. at this ledge, but in a short distance it changes. The shales are in general smooth, olive in tint, and in some that are a little mealy in texture are fossils, the most common species being *Pterochaenia fragilis* (Hall), *Buchiola retrostriata* (?) (v. Buch), *Goniatites*, *Coleolus*, etc.

This exposure is nearly all shale, and it belongs in the lower part of the Jennings formation. In fauna, lithologic appearance, and stratigraphic position it agrees closely with the Portage stage of the Upper Devonian of New York. It will be remembered that Dana united the Genesee, Portage, and Chemung to form his Chemung period, which corresponds very nearly, if not quite, to the Jennings formation.

On the bank of a run about one mile east of Moorefield Junction are olive, argillaceous shales by the side of the road, which alternate with thin, micaceous sandstones. No fossils were found, and the rocks are referred to the Portage stage. A dip of 45° S., 40° E. was noted. A ledge by the side of the road on the southern bank of Mill Creek, at the crossing directly east of Moorefield Junction is composed mostly of smooth, argillaceous, olive shales interstratified with thin sandstone layers which are rather micaceous. At the creek level is a sandstone layer two or more feet in thickness. The dip is about 2° west. No fossils were found and the rocks are referred to the Portage stage.

To the west of Moorefield Junction smooth, argillaceous, olive shales, alternating with layers of sandstone, occur along the highway and dip to the east. No fossils were found. These shales, as in the case of those east of Moorefield Junction, are considered as of Portage age and occur in the lower part of the Jennings formation.

Patterson Creek.—The next valley to the west of that of South Branch is that of Patterson Creek in Mineral County, and since in the volumes on the paleontology of New York certain Hamilton species are mentioned as occurring at Patterson's Creek, West Virginia, the outcrops in the lower part of the valley of this creek were examined. The Romney area of this valley is a continuation of the narrow band which lies between Collier and Nicholas moun-

tains, to the north of the Potomac River, in Allegany County, Maryland.

The railroad station and post-office now known as Patterson or Patterson Depot is the locality formerly known as Patterson Creek. There are no exposures in the immediate vicinity of the station, and the Baltimore and Ohio railroad track, along which are several cuts, was followed to the eastward. The first one east of the station shows massive Oriskany sandstone, the greater part of which is hard, light gray, and quartzitic, certain layers containing a considerable number of fossils, particularly specimens of *Spirifer arenosus* (Conrad), some of which are very perfect. At the eastern end of the cut the rocks dip sharply to the east, while farther to the west they are some distance above the track and nearly horizontal. In the second railroad cut east of the station is another outcrop of the Oriskany sandstone, which is also dipping to the eastward. In some of the partly decomposed layers of brown sandstone there are good specimens of *Spirifer murchisoni* Castelnau, associated with other Oriskany fossils. From near the center of the cut to the western end there is some of the very hard, quartzitic sandstone. The dip is about 23° S., 70° E.

At the first hill east of the cut described above are thin, bluish, arenaceous shales which form a low ridge. These shales are rather firm, and when crumbled the pieces are larger than those on the Williams Road to the north in Maryland. No fossils were found. These shales are apparently in the lower, fairly barren portion of the Hamilton member of the Romney formation. It is to be remembered in this connection that the lower part of the Hamilton formation consists of rather arenaceous shales, in which fossils are of infrequent occurrence. This area is the southwestern continuation of the belt of the Romney formation to the east of Collier Mountain in Allegany County, Maryland.

Along the hill by the highway about one mile south of Patterson Depot are outcrops of thin, bluish-black shales which are fairly argillaceous and split into thin pieces weathering to a brownish color. In certain layers fossils are rather thick, especially *Ambocoelia umbonata* (Conrad) and *Liorhynchus limitare* (Vanuxem). Specimens of *Coleolus* and small-winged Lamellibranchs were also

seen. On the ridge to the west is the Oriskany sandstone, dipping to the east, as shown in Rocky River and on the highway from the west perhaps one-sixth of a mile north of this locality. These shales contain the diagnostic species—*Liorhynchus limitare* (Vanuxem)—of the Marcellus shale and agree fairly well lithologically, as far as exposed, with those referred to the Marcellus shale in the Baltimore and Ohio railroad cut at the 21st Bridge, Maryland; consequently, they are referred to the Marcellus member of the Romney formation.

The valley to the east of this ridge along Patterson Creek is presumably composed of the Romney shale, which is probably in a small syncline. The steep wooded ridge to the east of Patterson Creek is the one in which occurs the first railroad cut east of Patterson Depot. The Patterson Creek Road was followed fully two miles from the station, and after the first exposure of the black Marcellus shales they are seen at frequent intervals at the roadside. The more southern shales examined, however, were rather coarser, somewhat more arenaceous, and less fossiliferous. The greater part of the Marcellus is slightly arenaceous and with less clay than is found in the exposures on the Williams Road to the northwest in Maryland, and the shales do not disintegrate so readily.

On Plum Run, two miles above Patterson Depot and below Mr. Robinson's house, are coarse, unfossiliferous, arenaceous shales varying to fairly thin-bedded sandstones of bluish-gray color. The zone, however, is clearly above the fine black shales of the Marcellus, which contain numerous specimens of *Liorhynchus limitare* (Vanuxem).

On Mr. Robinson's farm is a small stream to the southwest of Plum Run, on which are bluish shales and sandstones, some of the latter being very hard and slightly calcareous. It is probable that this sandstone zone corresponds with the lower one of the Hamilton member of the Romney formation on the Williams Road, Maryland. Specimens of *Spirifer mucronatus* (Conrad) and *Ambocoelia umbonata* (Conrad) occur sparingly. On the small hill to the south of the run are yellowish to greenish, very argillaceous shales, containing some fossils. *Chonetes* is abundant in certain thin layers while other species, occurring more sparingly, are *Spirifer mucro-*

natus (Conrad), *Spirifer audaculus* (Conrad), *Nuculites oblongatus* Conrad, *Phacops*, *Pleurotomaria itys* (?) Hall, *Orthoceras telamon* (?) Hall, and *Macrochilus hamiltoniae* Hall.

In color and composition these shales are fairly similar to those above the lower sandstone on the Williams Road, Maryland; but on weathering they seem to have hardened to some extent. This shale contains, here and there, numerous small hard concretions as well as those of clay-iron stone, and it is also blotched with spots and streaks of dark-red color from the weathering of the iron. It is very probable that the specimens of *Nuculites oblongatus* Conrad, *N. triqueter* Conrad, *Palaeoneilo constricta* (Conrad), *P. tenuistriata* Hall, *P. virginiana* Hall, *Grammysia alveata* (Conrad), *G. arcuata* (Conrad) listed by Hall from Patterson's Creek, Va.,[1] and perhaps other species, came from this portion of the Romney formation. These outcrops have a lithologic similarity to beds across the Potomac River in Maryland which have been correlated with the Hamilton and are their southern continuation. Furthermore, these rocks contain fossils that are common in typical outcrops of the New York Hamilton, and for these reasons they are referred to the Hamilton member of the Romney formation.

NOTE ON THE CORRELATION OF THE MARYLAND HELDERBERG FORMATION

Dr. Swartz in his "Introductory" account of "The Helderberg Formation" states that "Rowe divided the Helderberg into the equivalents of the Manlius, Coeymans, New Scotland, and Becraft formations of New York. O'Harra, who prepared a report upon the geology of Allegany County about the same time, assigned the same limits to it."[2] Mr. Richard B. Rowe was one of the writer's students when he was professor of geology in Union College, where Mr. Rowe studied under his direction the Helderberg formations in their typical region in the Helderberg Mountains of eastern New York. Later Mr. Rowe went as a graduate student to Johns Hopkins University and began the field study of the Paleozoic formations of western Maryland during the summer of 1897. In

[1] *Palaeontology*, Vol. V, Part I, Lamellibranchiata II, 1885.

[2] *Maryland Geological Survey, Lower Devonian*, text (1913), p. 97.

1898 the writer became Chief of Division of Appalachian Geology of the Maryland Geological Survey, and with Mr. Rowe as his principal assistant devoted that summer to field work on the Paleozoic formations of western Maryland. We worked together in the field and were fully agreed concerning the correlation of certain limestones with the Coeymans, New Scotland, and Becraft formations of New York, which is believed to be the first definite recognition of them in Maryland. A statement of the relationship of this work was published by the writer in 1901,[1] together with the correlation of the Helderberg limestone with the New York formations.[2] This was recognized by Mr. George W. Stose, who wrote as follows:

> The presence in this region [Pawpaw-Hancock Folio] of a representative of the Helderberg group of New York has long been known, but in earlier reports of the United States Geological Survey and of the Maryland Survey it was not differentiated as a separate formation but was included with the underlying calcareous formations under the name Lewistown. In 1901, Prosser, reporting the results of studies made by him and his associates for the Maryland Geological Survey, later embodied in the State report [Vol. VI (1906), 133, 134], recognized in the Maryland rocks the various faunal divisions of the Helderberg of New York.[3]

The writer in his *Historical Review and Bibliography* of the Maryland Devonian summarized the contents of his 1901 article on "The Paleozoic Formations of Allegany County, Maryland."[4] The notice of "The Geology of Allegany County" in the same review contains the statement that "The nomenclature and classification of the formations are those adopted by Messrs. Clark, Prosser, and Rowe for the Maryland formations."[5]

[1] *Journal of Geology,* IX, (1901) 409.

[2] *Ibid.*, 415, 416.

[3] *Geologic Atlas of the United States*, Folio No. 179 (1912), p. 9, col. 1.

[4] *Maryland Geological Survey, Lower Devonian,* text (1913), p. 51.

[5] *Loc. cit.*

THE STRENGTH OF THE EARTH'S CRUST

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PART VII. VARIATIONS OF STRENGTH WITH DEPTH AS SHOWN BY THE NATURE OF DEPARTURES FROM ISOSTASY

SECTION B, APPLICATIONS OF THE THEORY

GEODETIC EVIDENCE AS TO LIMITING HEIGHTS AND WAVE-LENGTHS

Measurements of strength by maximum loads.—In the course of geologic time the internal forces of igneous intrusion and tangential compression, the external forces of erosion and sedimentation, have tended to strain the crust to its limits of strength, and the degree of isostasy which exists constitutes a measure of those limits. Small loads of large wave-length and large loads of small wave-length will tend to rise to maxima which may be used in connection with the theory of distribution of stress, as considered in Part VII, Section A, to give an approximate idea of the limits and distribution of strength in the crust.

The problem is to find the maximum vertical load and its relation to wave-length acting over areas which may be regarded as

forming roughly a harmonic series. The theory applies best where elongated unit areas are flanked by areas of opposite sign. But even where a single axis of uncompensated elevation or depression is surrounded with a region of mean elevation it may be regarded as a half of a wave sustained by the rigidity of the earth. The stresses in the vertical plane through its crest line would appear to be less than, but not so greatly different from, what they would be for a continued series. Where the load is of oval instead of zonal distribution the stresses would also be somewhat diminished in case the oval is surrounded by a neutral region, but if a series of ovals of opposite signs is analogous to two intersecting wave-series, although the stresses would be complicated and are not in general the sum of the stresses due to the separate series, yet it does not appear that the extreme maxima would be necessarily less than the sum, and many of the maxima would be greater than the maxima of the separate series.

Harmonic loads with short wave-lengths.—Horizontal compression builds mountain folds of which the individual ranges are clearly the results of compression and not of isostatic elevation. Erosion dissects an elevated country on a pattern of a certain scale, deep valleys of erosion separating the crust remnants above base-level which are not yet consumed. These actions produce a surface relief which corresponds roughly to various harmonic series of appreciable amplitude and wave-length, but in this section will be considered only the geodetic evidence of variable mass not isostatically compensated, much of the variation being due to the concealed heterogeneities of density.

The distance between Washington, D.C., and Hoboken, New Jersey, as estimated in Part V, Section B, is 326 km. Within this distance are three intermediate stations and the two limiting stations, each station showing a gravity anomaly opposite in sign to that of the adjacent stations. There must be then at least two wave-lengths. The average change of anomaly between the adjacent stations is 0.021 dyne. As it is wholly improbable that the stations are located at the crest lines of the waves, the whole amplitude may safely be taken as at least 0.026 and the wave-length 160 km.

Seattle and Olympia are 80 km. apart and the difference of their anomalies is 0.126 dyne. If the anomalies had been measured at the points giving maximum values they would certainly give a difference of at least 0.130 or perhaps 0.140, about five times the amplitude of the variations between Washington and Hoboken. This large value seems, however, to be exceptional for the United States and may constitute but a single wave. We may take it, however, as showing that the crust can sustain a harmonic load of 160 km. (100 miles) wave-length and total amplitude of 0.120 dyne. For the reasons explained in Part IV, especially on p. 304, the divisor to be used to turn this anomaly into the equivalence of rock measured in feet could not be over 0.0018 and a better figure for the interpretation of this short wave-length is 0.0015. This gives an amplitude of 8,000 feet (2,440 m.). The part below the mean level is then 50 miles wide and 4,000 feet deep, the adjacent positive parts of the wave being of equal dimensions but in the opposite direction. This is of the same order of relief as the larger ranges of folded mountains and intermontane valleys. The stresses which this harmonic series imposes upon the crust are shown by curve A of Fig. 18.

Maximum loads for mean wave-lengths.—In Part II it was argued that the evidence of anomalies from mountain stations showed regional compensation on the average probably to the outer limit of zone O, radius 166.7 km., diameter consequently 333.4 km.

Over the United States in general the intercepts of the areas of grouped deflections averaged 180 miles. The average diameter would therefore doubtless be as great as 200 miles (320 km.).

On Bowie's map of the New Method gravity anomalies,[1] it is seen that the distances from pronounced maxima to pronounced minima average 250–350 km.

From these several lines of evidence we may conclude with some confidence that the half wave-length for pronounced anomalies in the United States averages near 300–350 km. The wave-length is therefore 600–700 km. (373–435 miles). A wave-length of 600 km. will be taken. The pronounced maxima for these waves runs from plus or minus 0.030 to 0.060 dyne. The real maxima

[1] This article, Part II, *Jour. Geol.*, XXII (1914), 153.

are to be regarded in most cases as situated to one side of the stations and somewhat greater. But exceptional and local maxima must be smoothed out to form a part of the harmonic curve. It is, furthermore, the difference of anomaly between adjacent opposite phases which is the significant feature. This difference runs from 0.060 to 0.080. The latter figure will be chosen. For this wave-length, representing a certain unit area of attraction, the best divisor is perhaps to take 0.0024 dyne of anomaly as equivalent to 100 feet of rock. An anomaly of 0.080 dyne is on that basis equivalent to 3,330 feet (1,015 m.) of rock. The crust of the United States sustains, therefore, harmonic loads 600 km. (373 miles) in wave-length and 1,015 m. (3,330 feet) in total amplitude. The stresses which this wave-series imposes on the crust are shown by curve B, Fig. 18.

Departures from isostasy of large wave-lengths.—For the continent as a whole and in its relations to the ocean basins isostasy is nearly perfect; but the question rises here, how nearly? The first term of the gravity formula for the Vienna system of gravity observations is 978.046 dynes. The first term for the Potsdam system is 978.030. The first term for the United States system after rejecting the Seattle anomalies is, as shown by Bowie, 978.038 dynes. These respective systems differ as a whole by these amounts. We have no right to assume that any one is absolutely correct. The whole of the United States system may lie a little above or below the level giving isostatic compensation with respect to the average surrounding ocean basins, or with respect to the entire earth. The mean value for the United States suggests, however, that, as a whole, the continent lies within a few hundred feet, possibly less than one hundred feet, of the level which would give perfect isostatic equilibrium.

Let us consider next its larger parts. These can be compared with each other and with the United States as a whole. Although, as discussed in Part IV, the map of gravity anomalies lacks detail, the grouping of many stations of like sign into large areas gives confidence in the conclusion that there are regional departures from isostasy. These are of two or three orders of magnitude, of which the areally smaller have been discussed. To bring out the

areally larger we must draw boundaries about large regions which show a dominance of anomalies of one sign. These boundaries, however, must be taken so as to give compact unit areas, so as not to obtain an unreal result by the political expedient of gerrymandering the districts.

Select as a center the point whose geographic co-ordinates are lat. 42°, long. 102°. Describe about this center a circle of 850 km. radius. This includes an area equal to 29 per cent of the area of the United States. It should be taken as including the negative anomaly station 99 on its southern border. This circle covers a large positive region which could be made still more positive by an extension of its boundaries to the northeast over Wisconsin and Michigan. Within this circle are distributed with a fair degree of uniformity 31 of the 122 gravity stations of the United States. The mean with regard to sign of the anomalies of these 31 stations referred to the United States mean with regard to sign is +0.010 dyne. As the mean without regard to sign of all stations in the United States excluding Seattle is only 0.018, it is seen that this positive region stands out clearly from the general average.

West of this circle and, on the south, to the west of long. 107° there are 21 stations, including one of the two Seattle stations. These mark a broad region of negative anomaly. The mean anomaly with regard to sign is −0.017 dyne. There seems to be no reason for completely omitting the exceptionally large Seattle anomalies. One of them has therefore been retained, but if both are omitted the mean is still −0.013. The value of −0.017 will here be adopted. The difference of the means of the central and western regions is consequently 0.027 dyne. Let these be regarded as the positive and negative phases of an harmonic wave and the mean departure of the two phases becomes 0.0135 from each side of the mean plane. Now it may be computed for a harmonic wave represented by the formula $y=A \sin Bx$ that the mean height of the wave above the mid-plane is 64 per cent of the crest height. From mid-plane to crest of this wave-series is therefore 0.021. From the large negative anomalies along the Pacific coast it would seem that this negative zone must extend somewhat further. The wave-length of this series is consequently between

2,600 and 3,000 km. A mean value of 2,800 km. (1,740 miles) will be chosen. From the breadth of half a wave-length it appears that 0.0034 dyne of anomaly may be taken as equivalent to 100 feet of rock. This gives the crest and trough as 625 feet (190 m.) from the mean plane, a total amplitude of 1,250 feet (380 m.). The stress-differences which this wave-series throws upon an earth elastically competent throughout to bear the stresses are shown by curve C, Fig. 18. Helmert has published an extensive paper dealing with the force of gravity and the distribution of mass in the crust of the earth,[1] to which the writer's attention has been called recently by Professor Pierpont, of the mathematical department of Yale University. In this paper Helmert adopts the hypothesis of regional isostasy and finds his results confirmatory of it, but not in accord with the hypothesis of close and local isostatic adjustment. His work is especially valuable as confirmatory of the present conclusions, since it deals with regions outside of the United States. As he does not, however, compute the corrections due to the distant large irregularities of topography, his figures cannot be directly compared with Hayford's New Method anomalies. Nevertheless his conclusions as to the existence of broad regional excesses or defects of mass are comparable to those here reached. Under the section on the horizontal displacement of compensation and extended excesses and defects of mass[2] he sums up part of the evidence in the following statement: "We have then to deal with a continuous region of positive total gravity disturbance in Europe 1,000 km. broad and also with a region of negative disturbance in Asia of at least 500 km. breadth, both possessing great linear extension."

RELATIONS OF ACTUAL STRESSES TO THE SUM OF HARMONIC WAVES

Both Darwin and Love point out that the actual stress-differences imposed by the superposition of different harmonic waves is not in general the sum of the individual stress-differences. Darwin, however, states the special conditions under which the

[1] "Die Schwerkraft und die Massenverteilung der Erde," *Ency. Math. Wissenschaft*, Band VI, 1, B, Heft 2 (1910), pp. 85–177.

[2] *Op. cit.*, pp. 152–54.

resultant is the sum of the individual stress-differences.[1] The three waves which have been considered are types which coexist and are superimposed. The total stress which they give would vary from

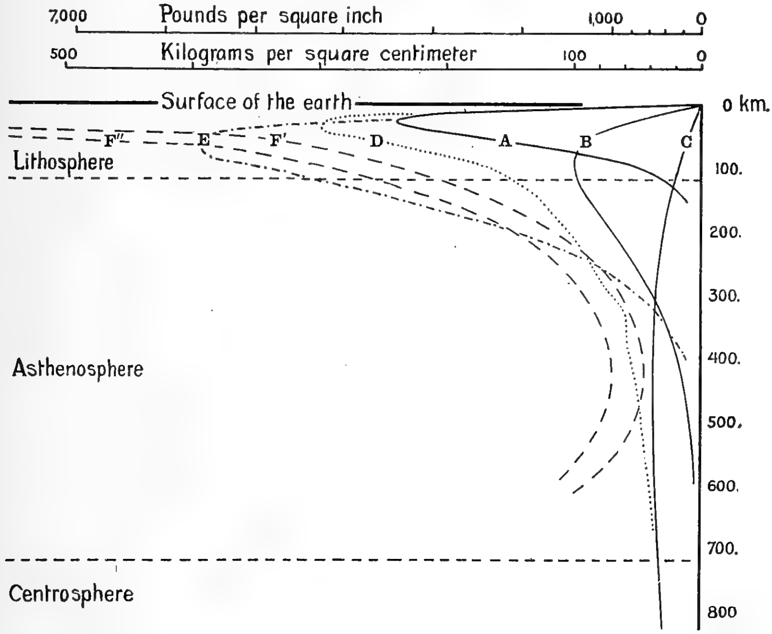

FIG. 18.—Stress-curves for harmonic waves on an earth elastically competent throughout, the waves representing departures from isostasy in the United States as given by analyzing the geodetic data into the following harmonic waves:

A, wave-length 160 km., amplitude 0.120 dyne=2,440 m. of rock.

B, wave-length 600 km., amplitude 0.080 dyne=1,015 m., of rock.

C, wave-length 2,800 km., amplitude 0.042 dyne=380 m. of rock.

D is the sum of A, B, and C.

E, wave-length 400 km., amplitude 0.366 dyne=4,000 m. of rock (from the Pacific Ocean).

F′, curve of strength suggested by geodetic evidence from the United States.

F″, curve of strength suggested by geologic evidence from various parts of the world.

complete neutralization up to their sum as a possible maximum. Curve D represents such an addition of A, B, and C, Fig. 18. There are reasons why this curve may be taken as a fairer representation of the maximum stress conditions under the United States than

[1] *Scientific Papers*, II, 492.

any individual curve, even though there may be no place where the culminating phases of like sign all coincide and become additive. These reasons are found in the subsurface location of those loads due to outstanding density and also to the added stresses due to isostatic compensation. These causes result in throwing a greater stress upon the outer parts of the lithosphere and also serve to broaden downward the stress diagrams. Furthermore, the stresses due to isostatic compensation of continents would appear to be in reality much greater under the margins than the small values computed by Love,[1] since he has taken the continents as having the broad sweeping surfaces of a harmonic nature, whereas, as a matter of fact, the continents slope off rather abruptly to the depths of the oceans. Facing the Pacific in fact, the two Americas show high mountain elevations. This would cause the stresses in the vicinity of these continental margins to resemble those imposed by a great mountain chain and its isostatic compensation rather than those imposed by the breadth of a continent. If isostatic compensation is complete under mountain slopes, Love shows for the cases computed by him that the maximum stress is about equal to that given by a column of rock one-fourth the total height from mountain crest to valley bottom. If the abyssal slopes of the continental platforms be taken as averaging 3–4 km. in elevation and 50–100 km. in width, it is seen that, even if fully compensated, they add stresses to the crust which may approach in magnitude one-half of the stresses shown by curve A of Fig. 18. The extreme depths of slope are much greater and it is clear that isostatic compensation cannot be exact under these great reliefs. Therefore we may conclude that curve D does not overestimate the maximum stresses imposed by the irregularities of the crust, both compensated and uncompensated, as indicated by geodetic evidence within the United States and especially along its ocean borders. This investigation, however, has been of a general nature and is designed merely to establish an order of magnitude. It remains for future work to make more precise analyses for each locality from the data which may be acquired, and especially to investigate quantitatively the problems offered by critical areas.

[1] Some Problems of Geodynamics (1911), chap. ii.

GEOLOGIC SUGGESTIONS AS TO MAGNITUDES OF CRUSTAL STRESSES

Submarine geanticlines and geosynclines:—Here will be considered some geologic illustrations of departures from isostasy, arranged in order of harmonic wave-length. They are to be compared with the results obtained from the study of the geodetic data. Most of the geologic evidence is merely suggestive, not conclusive, since diametrically opposite opinions are held as to the probability of the visible load being offset by an invisible compensation. As suggestions, however, they are none the less valuable, and point the way to needed geodetic observations.

The mountain folds advance from Asia over the floor of the Pacific Ocean, forming the system named by Suess the Oceanides. Mostly hidden beneath the ocean surface, they have been but little affected by erosion. Their ridges and deeps mark the greatest mountain reliefs of the globe. It is probable that here, if anywhere, tangential pressures have forced the crust into folds whose height combined with span is as great as the strength of the crust can endure. To what degree the elevations and depressions are compensated is, however, unknown, and the great arches are supported in part by the lateral pressure of the ocean water. It is quite possible if not probable that appreciable changes of deep-seated density may accompany the growth of such ridges, especially as they mostly exhibit a volcanic activity and are to a greater or less extent structures built up by igneous extrusion. It is not at all probable, however, that they are completely or possibly even largely compensated, but where the mountain folds and trough-like deeps broaden into plateaus or anti-plateaus the presumption is strengthened that the forms may there be isostatically compensated to a large degree. Such plateaus or anti-plateaus cannot then be used in the present argument. The ridges and troughs, however, show in their forms, as has been stated, modes of construction which are not conditioned on isostasy. Let attention be turned then to the folds of the ocean floor.

Passing from west to east, first may be noted between the Philippines, Borneo, and New Guinea a complex of ridges and basin-like deeps. The larger wave-length of that region runs from 300 to 500 km. The Ladrone Islands and Nero Deep give a

distance of about 150 km. from crest to trough and a wave-length of 400–500 km., these folds and many others exhibiting a lack of symmetry. The existence of strong folding pressures and a tendency to overthrusting and secondary vertical faulting seem to be expressed therein. The great fold passing north from New Zealand and showing as the Kermadec and Tonga islands with their fore-deeps gives a distance from crest to trough of 120–180 km., a wave-length of 400–500 km. Lower California and the troughs on each side show a wave-length of 350 km. The region of the Lesser Antilles is tectonically a northward branching of the Andean mountain system and shows, like the folds of the Pacific, crustal undulations with a wave-length of 350 km. We may conclude then that these folds of the ocean floor have a marked tendency to a wave-length of 300–500 km., there being commonly one great asymmetric fold passing out into subordinate marginal folds. The volcanic chain of the Hawaiian Islands shows, however, no related deeps and has a half wave-length of about 200 km.

Hayford and Bowie have given the New Method anomalies for a few stations in these regions.[1] A portion of the data has been abstracted and given in Part IV of the present article.[2] Four observations of Hecker's for the Tonga Plateau and Tonga Deep are given. They may not be of high value, since the method has been criticized as not possessing accuracy comparable to observations made by pendulum upon land. Furthermore, the four observations, two over the plateau and two over the deep, are spread through a distance of 5,100 km. along the axis of the structure instead of being taken on a transverse section. Nevertheless, as the reliefs and the corresponding anomalies are all of great magnitude, the errors become relatively small and the mean of the observations is therefore of some value. The two New Method anomalies for the Tonga Plateau give a mean of +0.202 dyne, the depth of water being 2,700 m. The two New Method anomalies for the Tonga Deep give a mean of −0.172 dyne, the depth of water averaging 7,500 m. If the amplitude

[1] *The Effect of Topography and Isostatic Compensation upon the Intensity of Gravity*, 1912, p. 81.

[2] *Jour. Geol.*, XXII (1914), 311.

of the uncompensated portion of the crust-waves be measured in terms of anomaly by taking the algebraic sum of the anomalies over the plateau and the deep, a total amplitude is obtained of 0.374 or a half-amplitude of 0.187. On the island of Hawaii an observation on Mauna Kea at an elevation of 3,981 m. gave a New Method anomaly of +0.183, almost the same figure as the half-amplitude for the great Tonga crust-wave.

Helmert has discussed the gravity disturbances found in the Hawaiian Islands and states of them: "For the Hawaiian Islands it must be concluded on the whole that a part of the mass gives rise to positive gravity disturbances and only the remainder is isostatically supported. If the disturbances were produced solely by the mass of the islands the values of Δg and $\Delta g''$ [the disturbances of gravity] would be somewhat greater than they are found."[1]

From this review of the mountain chains of the Pacific it may be concluded that the ocean floor can sustain a harmonic wave-length of 400 km. which gives an anomaly at the crest lines as great as that observed on Mauna Kea, 0.183 dyne. To interpret this as an equivalent load of rock a divisor must be selected. The divisor depends upon the depth and distribution of compensation and the area of the region of outstanding mass. As shown in Part IV, p. 311, 0.0024 dyne might reasonably be chosen as the amount of anomaly equivalent to 100 feet of rock, but for these great loads it is desirable to lean toward the side of an underestimate. Therefore 0.0030 will be taken as the divisor. This gives 1,868 m. as the crest height of the uncompensated portion of the Hawaiian mountain chain. The same applies to the Tonga fold. If, however, 0.0024 should be chosen as the divisor, then 0.183 dyne of anomaly would correspond to a half-amplitude of 2,334 m.

It may be taken then as fairly certain that these great mountain chains show reliefs which depart as much as 2,000 m. above and below the mean level which would give perfect isostasy. It may be concluded in consequence that the oceanic crust can sustain a harmonic wave-length of 400 km. with an uncompensated amplitude measured by 4,000 m. of rock. The diagram of

[1] "Die Schwerkraft und die Massenverteilung der Erde," *Ency. math. Wissenschaften*, Band VI, 1, B, Heft 2, (1910), p. 133.

stress-differences for this is shown in curve E, Fig. 18. But if the rock has a density of 2.67 and the sea-water a density of 1.03, this corresponds to an amplitude beneath the ocean surface of 6,513 m. of uncompensated rock. This is only about two-thirds of the maximum relief which is observed, so that it is well within the limits of possibility. These few available figures suggest that the sharp submarine ridges and deeps may not be more than one-third or two-thirds compensated.

The Niger delta.—Reverting to the discussion of the Niger delta given in Part I, it is seen that there is no evidence of depression around its margin. It may be taken then as the positive half of a harmonic wave well within the limits of crustal strength. If the section of the delta be taken as given in Figs. 3 and 4, pp. 31, 43, it is seen that the load is disk-like in form, instead of being indefinitely elongated at right angles to the section in accordance with the form of a zonal harmonic. It seems likely, because of these two departures from the nature of a harmonic series, that the stresses beneath it are not more than half of those which would be given by the completed harmonic curve. As it is merely the order of magnitude of the stress-differences which we may hope to attain we may proceed in accordance with these rough assumptions. It is seen that the section of the delta shows a half wave-length of about 300 km. and a maximum thickness equivalent to 1,650 m. of rock upon land. This corresponds to the half-amplitude or thickness above the mean plane. If half of this is taken as a measure of the stress, it gives a wave-length of 600 km. and a total amplitude of 1,650 m. The stress-curve for this harmonic series is 60 per cent larger than the stresses due to the outstanding masses of the same wave-length as given by the geodetic evidence in the United States and shown in curve B, Fig. 18. As the estimate from the Niger delta is very imperfect and unchecked by pendulum observations reduced by the New Method, the stress-curve is not plotted.

The existing continental ice sheets.—Two ice sheets of sub-continental proportions remain in existence, the Greenland and Antarctic. They form great plateaus sloping upward from the margins; the Greenland sheet reaching elevations at its center

between 9,000 and 10,000 feet, the Antarctic attaining to about 11,000 feet. The average thickness of the ice must be thousands of feet in each case. The development of these ice caps during the refrigeration of climate which marked the later Tertiary must have imposed upon the crust great loads of wide span. If isostatic equilibrium was previously complete to a large degree, the ice mantles should give valuable measures of crustal strength. For this purpose, however, a set of gravity measurements should be carried inland and reduced by the Hayfordian method. The facts that these two ice-mantled areas are both high plateaus, and that no other adjacent unglaciated land is of similar topographic character, suggest that these regions may be competent to carry great thicknesses of ice without isostatic yielding. There is no present basis, however, for making a quantitative estimate. It must be borne in mind, furthermore, that the ice mantle is only about one-third of the density of rock and that lofty mountains exist in both regions, showing that these lands would possess considerable mean elevation even without the presence of the ice. The effect of the difference of density between ice and rock may be appreciated by considering that an ice sheet 3,000 feet in thickness would possess the same mass as a layer of rock 1,000 feet thick. For isostasy to remain perfect after the development of this ice sheet, the crust would have to sink 1,000 feet, but the surface of the ice would still be 2,000 feet above the former level and give an appearance of load which would not in reality exist.

This is a problem meriting research for several reasons. A knowledge of the load which is sustained by these regions would show to what degree the warpings connected with the extinct Pleistocene ice sheets were mere elastic responses to load, to what degree they marked subcrustal plastic flow working toward isostasy. The results could be applied also to the problem as to how far from isostatic equilibrium a continent might come to lie as a result of continent-wide base-leveling in a period of geologic quiet. It seems not impossible that the stress-curve due to the portion of the glacial load which is elastically sustained would give stress-differences greater at a depth of 300–500 km. than those shown by curve C, Fig. 18. Such an investigation may then

be an essential factor in measuring the maximum strength of the lithosphere and more especially the asthenosphere.

Accordance of geologic with geodetic evidence.—The United States and its bordering ocean bottoms is a region of moderate reliefs as compared to the great folds of the ocean floor or of the continent of Eurasia. The geologic forces of folding and uplift have not worked here with their greatest intensity and the central and eastern half of the continent has been affected by the world-involving Cenozoic diastrophism to only a moderate degree. It is to be expected then that the greatest strains upon the crust, the maximum departures from isostasy, would not be found here. In accordance with this expectation it has been seen that by far the greatest New Method gravity anomalies are found in other regions and associated in most cases with the greater reliefs of the globe. The geologic evidence is in harmony; the amount of uncompensated relief, parallel to the geodetic evidence, is greatest for the lesser wave-lengths; but, throughout, the geologic evidence suggests that the actual burdens which can be borne by the crust, as found in regions of culminating stress, are appreciably greater than those detected by geodetic methods as existing in the region of the United States.

If, in some past ages, as during the Appalachian or Sierran revolutions, strains were generated in this continent as great as those found now in some other regions, it would appear that the slow changes of geologic time, of erosion and crustal readjustment, have partially eased the crust of its load. We may have, then, a variable crustal strength—a maximum strength exhibited during and following the crises of great diastrophism; another, lesser strength, which measures the loads which the crust without failing can bear through all of geologic time.

ADJUSTMENT OF LOADS TO THE DISTRIBUTION OF STRENGTH

It has been seen that the departures from flotational equilibrium may become very notable and are of greatest vertical magnitude for wave-lengths from 100 up to 400 km. The strains generated by these loads, if distributed through an elastic crust, consequently

reach maximum values at depths not exceeding 64 km. Is this because the earth shell below the zone of compensation is strong, but for some unrelated reason free from large stress-differences, or is there an absence of such stress-differences because this shell is too weak to bear them? If the latter is true, then the relations of amplitude to wave-length which have been developed in this chapter offer additional proofs of the reality of the existence of the asthenosphere.

The geologic evidence on the evolution of continental structures and elevations leads to the conclusion that the distribution of stress-differences must be in reality the result of the existence of a zone which cannot carry large distortional strains, as may be seen upon brief consideration.

The internal activities of igneous intrusion and of tangential compression do not in themselves work toward isostatic equilibrium, but merely toward accentuation of relief. Erosion and sedimentation, while tending to destroy this relief, are not agents tending to create, but to destroy, such isostatic relations as have developed. All of these activities work on a continental or subcontinental as well as on an orogenic scale, as seen in the Cenozoic history of the broad Cordilleran province, yet while the orogenic departures from isostasy are vertically very great, the continental departures are very moderate. For the latter there must be then some more narrowly limiting condition. This corresponds to the incapacity of a deep zone, the asthenosphere, to carry large stress-differences and the incapacity of the lithosphere in spite of its greater strength to act effectively after the fashion of a beam for loads of great span. The orogenic structures, on the other hand, give maximum stresses much nearer the surface, in the stronger lithosphere; because of their shorter wave-lengths they do not produce in it bending stress as in a loaded beam and affect comparatively little the deeper-seated asthenosphere.

If, then, it is known from the preceding theoretical considerations that the limits of strength of the lithosphere and asthenosphere determine the limits of the departures from isostasy, the analysis of the nature of these departures may lead in turn to a knowledge of the distribution of strength.

CHARACTER OF THE CURVE OF STRENGTH

In curve F′ of Fig. 18 is shown the nature of the curve of strength as suggested by the geodetic evidence from the United States. In curve F″ is shown the nature of the curve as suggested by the departures from isostasy exhibited by the great mountain axes and possibly by the continental ice sheets. These curves may be taken as showing the value of the elastic limit at various depths for permanent stresses. With varying geologic conditions, especially those connected with rising magmas and their emanations, the curve of strength must vary widely, and furthermore no very close parallelism of strength-curve and stress-curve is to be expected. These curves, therefore, are intended to bring out general relations; they are of qualitative, not quantitative value. The drawing of curve F″ somewhat inside of curve E means that below the point of maximum stress in E, as given for a homogeneous elastic earth, the stress is assumed as somewhat greater than the crust at those levels can sustain. Upon the development of this load plastic flow in these deeper levels would take place maintaining the stress within the strength curve for each level; the crust above would come to act to some extent as a bending plate, the stresses within it would increase, chiefly within the upper and lower portions. This added strain would compensate for the yielding below. For the reasons discussed previously, however, showing the structural weakness of the lithosphere as a beam, this action, it is thought, could not go very far, and, in consequence, the loads on the lithosphere are essentially such as to give stresses contained within it, distributed according to Darwin's law. The preceding deals only with that part of the curve of strength which marks the gradation from lithosphere to centrosphere. The relations of this part to those above and below need still to be considered.

The highest stress found for the loads regarded as harmonic waves was for the great folds on the floor of the Pacific Ocean. These were taken as equivalent to harmonic waves of rock of density 2.67, 400 km. in wave-length, and 4,000 m. in amplitude. But even these folds give a maximum stress of only 393 kg. per sq. cm. (5,590 pounds per square inch), and this at a depth of 64 km. At the surface strong limestone or granite can

sustain a stress-difference of 1,750 kg. per sq. cm. (25,000 pounds per square inch), and selected specimens show ultimate breaking strength approaching 2,800 kg. per sq. cm. For stresses of geologic endurance and in the heterogeneous outer crust it is probable, however, that stress limits should be chosen below 1,750 kg. per sq. cm.

The work of Adams and King has shown that small cavities in granite are not closed when the rock is subjected to the pressure and temperature normally existing in the earth at a depth of 11 miles.[1] The presence of occluded gases acting through great lengths of time, by facilitating recrystallization, might affect this result of laboratory experiments, but the capacity of dry rock to sustain even greater cubic pressures without yielding seems to make safe the conclusion that except in the presence of magmatic emanations the crust at a depth of 11 miles (17.7 km.) is able to bear a stress difference of 100,000 pounds per square inch and is at least four times as strong as rock close to the surface.

At twice this depth, however, the temperatures become such that if it were not for the great pressures even dry rocks would approach a molten condition. The presence of high temperatures and of gases which may act as crystallizers presumably becomes dominant at such depths over the effects of the increasing pressures. We many conclude, therefore, that the maximum strength of the crust in regions free from igneous activity is found at levels above rather than below 40 km. and may lie between 20 and 30 km. deep.

To bring to a focus this discussion a tabulation of ratios of strengths for increasing depths may be given, as derived from the strength curves F′, F″ of Fig. 18, the standard being taken as the strength of surface rocks. By giving them merely as ratios and stating that the average strength of the solid rocks at the surface is itself an uncertain quantity owing to complications of structure and composition, the appearance of an undue certainty is avoided.

The general conclusion which stands out from this tabulation is that the weakest part of the asthenosphere is of the order of one one-hundredth of the maximum strength of the lithosphere and is perhaps only a twenty-fifth of that of massive surface rocks. Its

[1] *Jour. Geol.*, XX (1912), 97–138.

limit of capacity for sustaining stress-differences is apparently of the order of 1,000 pounds per square inch, though its weakness may be masked to some extent by the strength above. From the evidence, however, it seems capable of carrying stresses of more than 100 pounds per square inch, but is clearly incapable of carrying stresses of as much as 5,000 pounds per square inch. To reach a

TABLE XXX

ESTIMATED APPROXIMATE RATIOS GIVING THE VARIATION OF STRENGTH WITH DEPTH AS SHOWN BY THE NATURE OF DEPARTURES FROM ISOSTASY

LITHOSPHERE

Depth in Kilometers	Strength in Percentage
0	100
20	400
25	500
30	400
50	25
100	17

ASTHENOSPHERE

Depth in Kilometers	Strength in Percentage
200	8
300	5
400	4

more definite conclusion the subject must be tested from many angles and is a problem for the geophysicist rather than for the geologist, but the results are of geological importance and the geologic and geodetic data may turn out to have more determinative value on the distribution of strength than the evidence from tides and earthquakes.

[*To be continued*]

SOME ELLIPSOIDAL LAVAS ON PRINCE WILLIAM SOUND, ALASKA[1]

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Many descriptions of ellipsoidal, spheroidal, or pillow lavas have been published from time to time. Rocks of this type have a wide geographic range and have been formed at intervals from pre-Cambrian to the present. Various conclusions have been reached by different writers as to the mode of origin of ellipsoidal lavas, and the reasons for the development of their peculiar ellipsoidal, spheroidal, or pillow-shaped forms. The ellipsoids or pillows have been considered to be (1) concretionary, (2) the result of brecciation, (3) due to contraction, (4) accumulations of volcanic bombs, (5) the result of explosive eruptions, (6) and the result of submarine cooling. In a recent article N. Sundius[2] has well summarized the important literature on the subject.

Notwithstanding the various hypotheses which have been advanced to explain this phenomenon, it has come to be generally accepted that the ellipsoids are the result of the flowing of the lavas into water, or their extrusion under water, and are due to the rapid cooling of the flowing molten material under these conditions. Tempest Anderson[3] has described recent pillow lavas actually seen in the process of formation, and there now seems to be little reason to doubt that the chilling of lavas under water is responsible for many of the ellipsoidal lavas with which we have become familiar. The purpose of this paper is to present certain facts which seem to the writer to prove conclusively that the ellipsoidal lavas near Ellamar, Alaska, were subaqueous flows, and to give certain criteria

[1] Published by permission of the Director, United States Geological Survey.

[2] N. Sundius, "Pillow-Lava from the Kiruna District," *Geol. fören. Förhandl.*, Bd. 34, Häft 3 (1912), pp. 317–33.

[3] Tempest Anderson, "The Volcano of Matavanu in Savii," *Quarterly Jour. Geol. Soc.*, XLVI, No. 264 (1910), 621–39.

which, in this field at least, proved of value in determining which was the upper and which the lower side of a given flow as deposited, thus aiding in deciphering the structure in a region where the normal attitude of the beds had in many places been obscured, and where overturning of the beds might be expected.

The ellipsoidal greenstones here described were studied in the vicinity of Ellamar, Alaska, in the fall of 1912, during an investiga-

Fig. 1.—Cross-section of lava flow, showing ellipsoidal forms. The ellipsoids are characterized by both radial and concentric sets of cracks. Photo. by S. Paige.

tion of the geology and mineral resources of the district by Mr. B. L. Johnson and the writer. The exact age of the greenstones is not known, but they are associated with a thick series of sedimentary beds which are believed to be for the most part of Mesozoic age. The greenstones were of especial interest because all of the copper deposits of the district are in the greenstone, or closely associated with it. In those areas in which the greenstones are most highly mineralized the lavas have been extensively sheared and meta-

morphosed, so that they have almost completely lost their original appearance. In other places, however, the lavas have been tilted, with only slight folding and with little shearing, so that their original character is well preserved. The term greenstones has been commonly applied to these rocks, which are all diabases or basalts. The writer prefers the term "ellipsoidal" to other commonly used terms such as "spheroidal" or "pillow" lavas, for the forms are

FIG. 2.—Uptilted ellipsoidal flows, showing irregular shapes of the flow forms. Photo. by B. L. Johnson.

most often ellipsoidal in cross-section (Fig. 1), but in plan they appear as more or less continuous bodies with irregular bulges and protuberances (Figs. 2 and 3).

The best exposures for studying the ellipsoidal lavas were found along the shore from two to four miles northwest of the village of Ellamar, toward Rocky Point. There the waves have cut steep sea cliffs which rise from the water's edge, often with no beach at all, and the exposures are fresh and perfectly free from covering.

An examination of this section shows that although the ellipsoidal flows form the larger proportion of the series, they are interbedded with some more massive diabase beds. At intervals, however, the extrusive rocks are seen to be interrupted by slates and graywackes, the sedimentary beds becoming thicker and more frequent as the top of the greenstone series is approached. The strike and dip of adjoining lavas and sediments is the same, and it

FIG. 3.—Ellipsoidal lavas. The forms are ellipsoidal in cross-section, but in plan are long uneven bodies characterized by bulbous portions separated by constrictions. Photo. by S. Paige.

is evident that clastic sediments were deposited at intervals between the outpouring of the lava flows (Fig. 4). In a large number of observations on slate or slate-graywacke beds interbedded with ellipsoidal greenstones, no single case was found in which there was a discordance in bedding between the sediments and the lavas. Furthermore, in the cracks between the ellipsoids of the lower part of a flow and in the smaller cooling cracks within the individual ellipsoids there were frequently seen thin vein-like layers of a black, shaly material which proved to be of the same character as the under-

lying sediments. The presence of this sedimentary material in the cracks of the fractured ellipsoids can be satisfactorily explained only by assuming that the bed of mud upon which the lava was extruded was at that time soft and plastic, and that upon the cooling of the lavas, and their cracking from shrinkage as they chilled, the weight of the lava bed forced the soft mud from below up into the cracks of the lava as fast as they were opened. In places these

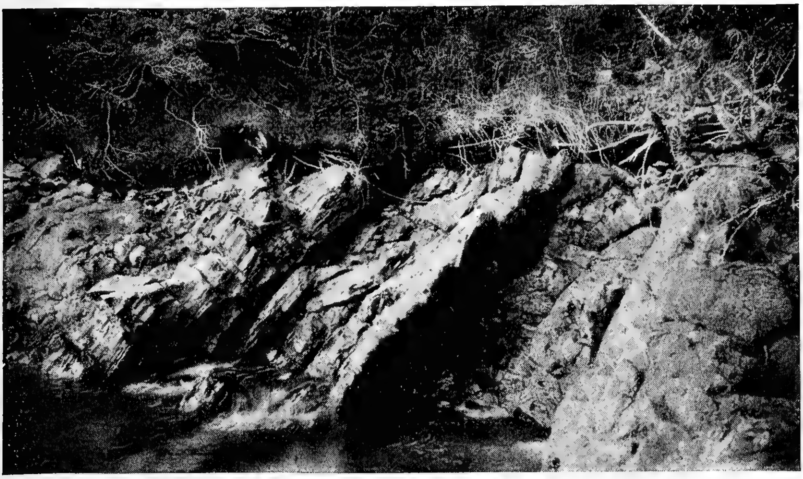

Fig. 4.—Steeply dipping ellipsoidal lava, on right, succeeded by conglomerate and graywacke. The uneven lower surface of the conglomerate conforms to the top of the underlying lava flow.

mud-filled cracks extend several feet into the flow, and where best developed give the rock a mosaic appearance, the fragments of greenstone, often only a few inches in diameter, being surrounded by a black layer of baked shale. The mud fillings of the cracks vary in width from the thickness of a knife blade to several inches. At one locality the openings between the ellipsoids were found to be filled with limestone.

In those places where thin ellipsoidal greenstone beds were adjoined both above and below by sedimentary beds, it was found that the two surfaces of the flow presented different appearances (Figs. 4 and 5). The under surface in each case was flat, the ellipsoids of the lower layer being flattened at the base, though presenting

curved surfaces on their other sides. At one place a steep overhanging sea cliff 30 feet high was formed by the base of such a flow, the underlying slate having been cut away. The base of this flow was flat and smooth, though the flow, as seen in section, was composed altogether of ellipsoidal lavas. Into the cracks in the base of this flow the mud had penetrated deeply. It is believed that the flat lower surface of this ellipsoidal flow could have been

FIG. 5.—Ellipsoidal lavas below, separated from more massive diabase above by an intercalated shale bed. The lower surface of the shale is uneven, having taken the shape of the ellipsoidal lava surface upon which it was deposited. The overlying flow is flat on the bottom, conforming to the shale surface. Photo. by S. Paige.

formed only by lava which was poured out upon a flat bed of sediments, the lava having conformed to the surface upon which it was deposited. The upper surface of an ellipsoidal bed presents a different appearance (Figs. 4 and 5). There the flow surface was under the pressure only of the column of water above it, and the surface hardened to form an irregular floor consisting of a succession of domes, like the surface of a magnified cobblestone pavement, the surface of the ellipsoids representing the cobble-

stones. Upon this surface later deposits of mud were laid down, the bottom of the mud bed taking the shape of the irregular lava surface. The lava flows show no signs of weathering or erosion upon their surface, and the succeeding mud beds must have been laid down soon after the lavas cooled.

Summary.—The greenstones near Ellamar, Alaska, comprise a thick series of lava flows, most of the flows consisting of ellipsoidal lavas, and many of them displaying this character in a striking way. At frequently recurring intervals in the greenstone series there are interbeded slates and graywackes, and rarely some limestone, the sedimentary beds being in every case structurally conformable with the overlying and underlying lavas. The surfaces of the flows show no evidence of erosion or weathering, the succeeding sediments having been laid down upon the fresh, unaltered lava surface. If the greenstones are subaqueous flows, as is believed to be the case, then they were merely interruptions during the deposition of ordinary clastic sediments.

If, on the other hand, the lavas were not subaqueous flows, and were extruded upon land, it is necessary to postulate an elevation of the surface above sea level for each extrusion of lava and an immediate resubmergence for the deposition of the sedimentary beds. In view of the many recurring sedimentary beds throughout the greenstone series, this latter assumption seems most unlikely. The absence of evidence of erosion or weathering on the surface of the lava flows also supports the conclusion that they solidified below sea level, as does the presence in the lava cracks of thin films of mud which must have been plastic at the time they were injected.

The difference in appearance of the flat bottom of an ellipsoidal flow, with its abundance of mud-filled cracks, from the uneven upper surface of the same flow, and the consequent unevenness of the bottom of the succeeding sedimentary bed often made it possible to determine which surface of a steeply tilted bed was originally the upper surface, and was a valuable aid in working out the structure of the beds.

A SKETCH OF THE LATE TERTIARY HISTORY OF THE UPPER MISSOURI RIVER

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The story of the Missouri River forms one of the most interesting chapters in the geologic history of the Great Plains. The valley of the present stream is composed of many parts, some relatively young and others moderately old, but as a unit it is youthful when compared with the valley of the Mississippi. Notwithstanding its youth, its history is so complex and so little known that only a mere sketch can be given at this time. All that will be attempted, therefore, will be to point out a few late Tertiary river channels, which may indicate former courses of the Missouri River and some of its larger tributaries, and sketch their bearing on the history of the main stream. One of these stream courses has been traced from Poplar, Montana, northeastward to the North Dakota boundary. Another valley, partially abandoned, extends northward from a point 6 miles west of the mouth of Bowlin Creek on the Little Missouri River to the mouth of Tobacco Garden Creek on Missouri River (see map, Fig. 1).

F. H. H. Calhoun[1] has mapped a number of preglacial valleys in north-central Montana, many of which are connected with the history of the Upper Missouri. One of the most important of these valleys extends from the mouth of Little Sandy Creek northeastward for about 35 miles to Milk River near Havre. Below its junction with this old valley the valley of Milk River is 2–2½ miles broader than above and clearly indicates that the headwaters of a larger stream, probably the Missouri, emptied into it and followed its course previous to the ice invasion. East of the mouth of Little Sandy Creek the present Missouri flows in a narrow, rocky channel for many miles and joins the old channel at the

[1] F. H. H. Calhoun, "The Montana Lobe of the Keewatin Ice Sheet," *U.S. Geol. Survey, Prof. Paper 50*.

mouth of Milk River, about 20 miles below Glasgow. Another old valley, noted by the writer in 1911, extends from Poplar, Montana, northeastward to Medicine Lake, thence eastward for several miles, thence northward to the vicinity of Dagmar and Coalridge, and thence northeastward to the lake region of the extreme northwestern corner of North Dakota. Beyond this the old river channel is completely buried by drift.

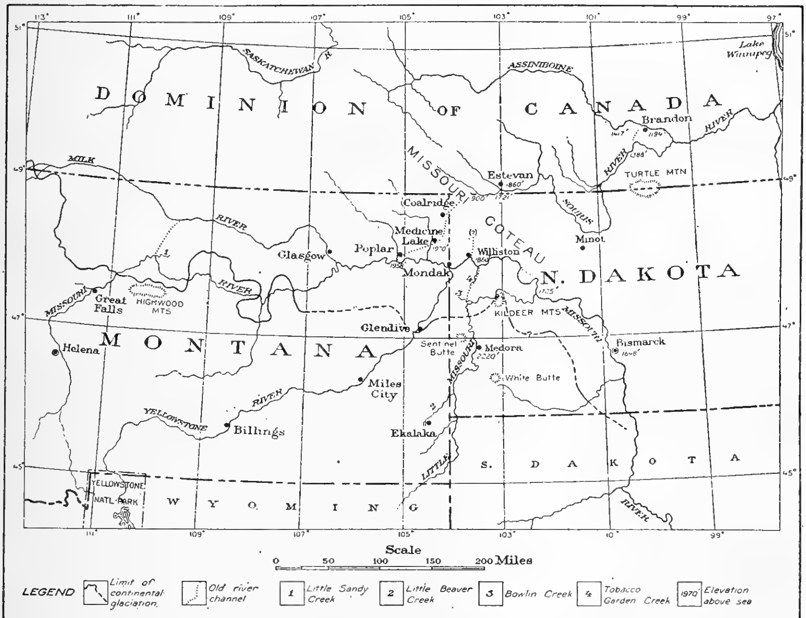

FIG. 1.—Map of Upper Missouri River, showing some Tertiary river channels. By C. M. Bauer.

In the vicinity of Poplar this old valley is nearly filled with glacial drift. The town of Poplar is built on a moraine, as shown by the section of drift 45 feet thick just north of the Indian School on Poplar River. Northeast of Poplar for several miles the drift is probably much thicker. In T. 28 N., R. 52 E., and northeastward for 20 miles the trough of the old valley is from 3 to 4 miles wide, with broad, gentle slopes on either side. Near Medicine Lake it widens, and on its south bank there is an area of sand dunes.

Small lakes are common along the old channel, and in the vicinity of Coalridge the valley trough is very distinct. A. L. Beekley traced this old valley from Medicine Lake to the North Dakota boundary in 1910, and describes its features in *Bulletin 471*, U.S. Geological Survey.

From the Missouri Coteau northeastward the ancient stream course is unknown, but it seems probable that it joined the preglacial Souris River and flowed thence into the predecessor of Assiniboine River, thence northward by way of Lake Winnipeg and Nelson River to Hudson Bay.

Elevations above sea along this course, some of which are shown on the map, point to this conclusion. Most of the altitudes given are of the lowest point in the immediate locality. If we take these altitudes in order, beginning at Poplar, Montana, and continuing northeastward, we find that, with the exception of Medicine Lake, which is 1,970, there is a gradual decrease in altitude from 1,958 feet on the present Missouri River bottom to 1,194 feet at Brandon on the Assiniboine River. It is significant that the present fall of the Missouri River in an equal distance is much less. The fact that Medicine Lake is now 12 feet higher than the flood plain at Poplar is probably due to postglacial erosion by the Missouri River. Another factor which must be considered is that the altitudes given do not indicate definitely the position of bedrock because of the varying thickness of glacial drift and alluvium in the old channel. It is also possible that differential warping to the amount of 200 feet, known to have occurred between latitudes 47° and 51° N., in the area occupied by Lake Agassiz and Lake Souris affected this region also and has relatively raised the country to the north from what it was in late Tertiary time.[1]

Changes in Yellowstone River due to glaciation are apparently few. Its present course, to within a few miles of its mouth, is about the same as it was in preglacial time. However, its anastomosing channel from Glendive to Mondak, on a flat, sandy, valley floor, is evidence of glacial filling in this portion of its channel. From a few miles south of Mondak the Tertiary Yellowstone probably fol-

[1] Warren Upham, "The Glacial Lake Agassiz," *U.S. Geol. Survey, Mon. 25*, pp. 474–522, 1896.

lowed the present course of the Missouri River to Williston, thence northward along the valley of Muddy Creek, which is much too wide for the stream that now occupies it. The writer has not traced this portion and was unable to get definite information on the old valley of the Yellowstone more than a few miles north of Williston.[1] However, Lambert, in *Exploration for the U.S. Pacific Railroad in 1853,* mentions a wide valley, discovered by Lander, which crosses the Coteau connecting the Souris with the Missouri near old Fort Union (Williston?) It seems probable, therefore, that the ancient stream proceeded north or northeastward from this point.

Concerning the course of the Tertiary ancestor of the Little Missouri River, the evidence is very plain. An old valley, several miles in width, extends northward from the first prominent eastward bend of the present stream near the mouth of Bowlin Creek to the head of Tobacco Garden Creek and thence along this valley to the present Missouri and probably joins the Tertiary Yellowstone near Williston. This old valley south of the Missouri River was first noted by F. A. Wilder,[2] who describes it as being evidently a former course of a prominent stream similar to the Little Missouri.

The present valley of the Little Missouri River from the Kildeer Mountains east to its mouth is narrow and bordered by rugged badlands of soft Fort Union strata. Consolidated glacial drift occurs on the jagged tops of some of the ridges 250 feet high and within one-half mile of the stream channel, but nowhere on the sides of the valley has consolidated drift been found. Although there are no terraces in this portion of the valley, in the vicinity of Medora and continuing southward into South Dakota a well-defined terrace from a few feet to several miles in width has been mapped by C. J. Hares.[3] Along Little Beaver Creek, one of the tributaries of the Little Missouri, the remaining fragments of this

[1] F. H. H. Calhoun, "The Montana Lobe of the Keewatin Ice Sheet," *U.S. Geol. Survey, Prof. Paper 50*, pp. 35, 36, 1906.

[2] F. A. Wilder, "The Lignite of North Dakota and Its Relation to Irrigation," *U.S. Geol. Survey, Water Supply, Paper 117*, p. 43, 1905.

[3] C. J. Hares, "Lignite in Southwest North Dakota," *U.S. Geol. Survey Bull.* (in preparation).

terrace have been mapped by the writer nearly to its headwaters. A few miles west of Ekalaka, Montana, several flat-topped, gravel-covered ridges 50 to 60 feet high form a prominent feature of the landscape. The terrace, where noted, is covered with gravel from a few inches to 10 or 12 feet in thickness. The gravel consists principally of red and white quartzite pebbles as well as some of chert and argillite, ranging from the size of sand grains to 2 or 3 inches in diameter. The terrace has its best development along the Little Missouri in southwestern North Dakota, and at this place it probably represents the late Tertiary valley floor of the Little Missouri, which existed until the invasion of the earlier ice sheet. Since then the Little Missouri has cut a channel to the east and shortened its course. The present trough of the Little Missouri, which is about 200 feet deep near Medora, is therefore the result of erosion since the disappearance of the early ice sheet.

Parts of the Missouri River above Poplar, Montana, and likewise the Yellowstone River, probably had their beginning at the end of the Cretaceous period, when the seas withdrew and exposed large areas in the region of the Rocky Mountains to stream erosion. These primitive streams and their associates were apparently small at first, but as the land area grew by withdrawal of the sea and with the rising of the mountain belt, in the early part of the Eocene epoch they became larger, gained velocity, and gathered sediment which was gradually spread out over the region now known as the great Fort Union area. These rivers doubtless meandered widely over this broad palustral flat and had their *embouchure* somewhere to the north,[1] probably into the predecessor of Hudson Bay or the Arctic Ocean.

As the primitive Yellowstone and Upper Missouri and their tributaries and allies continued their work, several thousand feet of sediment were deposited during the Eocene epoch in eastern Montana, western North Dakota, and parts of adjoining states. This was finally interrupted by an epeirogenic movement, which

[1] W. G. Tight, Abstract of paper read before the Geological Society of America, *Geol. Soc. Amer. Bull.*, XVIII.

Warren Upham, "Age of Missouri River," *American Geologist*, XXXIV (1904) 80–87.

caused a relative rise of this portion of the continent, together with some folding. The streams were now set for the task of erosion. During this time the Yellowstone succeeded in maintaining its course across the Glendive anticline, a long, narrow arch extending southeastward from the Yellowstone River, Montana, to the northwestern corner of South Dakota, and the Upper Missouri maintained its course across the Poplar dome, a low quaquaversal structure at Poplar, Montana.

Deposition again began on the Great Plains in Oligocene time.[1] The change in level causing deposition may have been the result of warping, the mountains on the west rising relative to the plains and furnishing more sediment.

A general uplift following the Oligocene inaugurated a great period of erosion and base-leveling on the northern Great Plains. In places, particularly in Saskatchewan,[2] the great rivers deposited some gravel and coarse sediment where the gradient became low and the load too great. However, during the late Tertiary, the Upper Missouri and Yellowstone carried away thousands of feet of strata and carved for themselves magnificent broad valleys in the soft Cretaceous and early Tertiary deposits.[3] In the vicinity of Highwood Mountain the erosion in the valleys probably amounted to about 3,000 feet. Farther east, near Sentinel Butte and the Kildeer Mountains, about 1,000 feet and near Turtle Mountain about 500 feet of material was excavated. Between these monadnocks the country was reduced to a peneplain. Then, according to Upham, followed the early Quaternary uplift and the glaciation of the northern part of the continent. The outlets to the north became blocked with ice and later were partly filled with drift and outwash. Ponding of waters and marginal drainage began. Lake Agassiz and Lake Souris were formed, as well as many other smaller lakes, most of which are now extinct. Just where the

[1] A. G. Leonard, geologic map of North Dakota, 1913, showing present distribution of White River formation.

[2] R. G. McConnell, "On the Cyprus Hills, Wood Mountain and Adjacent Territory," *Canada Geol. Survey Report*, N.S., I (1885) 70–c.

[3] Warren Upham, "Tertiary and Early Quaternary Base-leveling in Minnesota, Manitoba, etc.," *American Geologist*, XIV (1894), 235–46.

marginal drainage proceeded when the ice was at its maximum in eastern Montana and western North Dakota is at present a matter of conjecture. However, it is probable that it flowed southeastward across cols and through various preglacial valleys until it joined the Mississippi.[1]

After the retreat of the ice, the Upper Missouri, the Yellowstone, and the Little Missouri, owing to the glacial débris, no longer found an outlet to the north, but instead followed in part the glacial drainage and in part the preglacial stream courses southeastward to the Gulf of Mexico.

[1] J. E. Todd, "The Pleistocene History of Missouri River," *Am. Assoc. Adv. Sci. Proc.*, N.S., XXXIX (1914), 263–74.

THE INTERGLACIAL GORGES OF SIX MILE CREEK AT ITHACA, NEW YORK

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University of Illinois

INTRODUCTION

The question of the complexity of the glacial period in central New York is still largely unsettled. It is conceded probably by all that the region has suffered more than a single ice invasion, but how many such invasions there were and how far the different invasions were separated from each other in time are problems as yet only partially solved.

The criteria for the identification of distinct glacial epochs which can be applied in a hilly region well within the southern limits of the ice, and even within the zone of active erosion by the latest Wisconsin ice sheet, are necessarily different from those which have been used so extensively and so successfully in the Mississippi Valley. Under the conditions prevailing in central New York one could not expect, for instance, to find drift sheets imbricated in the same manner as on the plains of the Mississippi near the limit of glaciation where the erosive power of the ice was doubtless slight and undirected and where deposition was the rule. Nor would it be possible to apply the common test of the relative erosion suffered by drift sheets of different ages, for, since the region in question lies within the great moraines of the Late Wisconsin epoch, any erosional features developed on earlier drift sheets would have been modified or destroyed by the Wisconsin ice. The hilly nature of the country, too, would tend to prevent the accumulation of till in even, continuous sheets and would lead to an irregular, patchy distribution of any remnants of earlier drift deposits. The meagerness, in this region, of the evidences from glacial deposits indicating multiple glaciation may be judged from the fact that, so far as the writer has been able to discover, only one investigator, Carney,[1] has

[1] "Pre-Wisconsin Drift in the Finger Lake Region of New York," *Jour. Geol.*, XV (1907), 571–85.

definitely claimed to have found such deposits. Tarr[1] mentions two or three exposures of possible older drift near Ithaca, but believes that the evidence is inconclusive, and Gilbert[2] recognizes the possibility of two till sheets and the certainty of "an epoch of local till erosion by a glacier," but is led to no definite conclusions further than this.

There is, however, another criterion of multiple glaciation for the application of which the Finger Lake valleys furnish ideal conditions. This criterion is the amount of work performed during interglacial intervals by streams whose normal profiles have been thrown out of harmony by glacial erosion or deposition; streams which must, therefore, resort to gorge-cutting in their endeavor to bring their beds back to normal grade.

A glacier moving along a valley tends to overdeepen it and to leave tributary valleys hanging. As soon as the ice withdraws, the streams in the hanging tributary valleys begin the work of bringing their beds down to the normal profile of equilibrium. In doing this they cut gorges in the lower ends of the hanging valleys. Given sufficient time after an ice invasion, the tributary streams would enter the main valley at grade through narrow, gorgelike valleys cut in the bottoms of the hanging valleys. These gorges, narrow at first, would, after their streams have reached grade, continue to widen in a normal manner under the influence of the ordinary weathering agencies combined with lateral swinging of the stream.

After a single glacial epoch the streams in all the hanging valleys would be flowing in such gorges and would, in time, bring their beds down to a grade accordant with that of the main stream. Suppose, now, a second glacial epoch should intervene. The chances are that the main valley would be deepened still further by ice erosion while the gorges in the tributary valleys would become more or less clogged with drift. On the withdrawal of this second ice sheet the gorge bottoms of the tributary valleys would be left hanging above the bottom of the recently deepened main valley.

[1] Watkins Glen-Catatonk Folio, No. 169, U.S. Geol. Survey.

[2] G. K. Gilbert, "Boulder Pavement at Wilson, N.Y.," *Jour. Geol.*, VI (1898), 771–75.

Should a tributary gorge have been incompletely filled with drift, its stream would begin the task of clearing out the drift and of cutting a second gorge in the bottom of the earlier one. Should it happen, however, that the drift filling of a gorge was so complete that the stream on reoccupying its old valley and taking a consequent course found lower ground to one side, a new and independent gorge would be cut. Should the consequent course induced by the drift topography correspond in part with that of the old gorge and differ from it in part, the stream would cut a new gorge where it found itself out of the old channel, and merely clear out the old gorge where the two courses were in coincidence.

Parts of the old drift-filled gorge unoccupied by the stream would be preserved and would remain indefinitely as fossil gorges hanging above the level of the main valley bottom.

If an interval of deglaciation sufficiently long ensued, the tributary streams would again bring their beds down to grade. A third epoch of glaciation would bring about a repetition of events with the possible formation of still another gorge in the hanging valley bottom. The number of distinct gorges should, then, give at least the minimum number of glacial invasions which the region had suffered. It would not necessarily give the maximum, for during one or more of the intervals the stream might have merely re-excavated one of the older gorges.

The relative width of gorges cut during different intervals should, other things being equal, give a rough measure of the relative length of the interglacial intervals which they represent. A comparison of the width of the older gorges with that of the postglacial gorge should, in the same way, give a rough measure of the length of the interglacial interval as compared with postglacial time. This would be true, of course, only providing the older gorges were occupied by streams during only one interval.

The criterion outlined above for the determination of repeated ice invasions is considered to have advantages over others in that the possibility of minor halts and readvances of the glacier being interpreted as distinct glacial epochs is reduced to a minimum, for streams require time for the excavation of gorges. If an interglacial gorge is as large or larger than the postglacial gorge of the same

stream, it is a fair inference that the interglacial interval bore a similar ratio to postglacial time. It would be necessary, however, to take account of possible different gradients of the streams and different climatic conditions during the intervals, but the effects of these could hardly lead to the confusion of true interglacial intervals with minor retreats of a fluctuating ice front.

The conditions outlined in the preceding paragraphs are realized very fully in the Finger Lake valleys. The main troughs, extending approximately in the direction of ice movement, were deeply scoured, while tributary valleys, suffering less erosion, were left hanging. The extent to which hanging valleys were developed has been so well presented in the writings of Tarr and others that it need not be dwelt upon here.

If the glacial period in this region was represented by more than one ice invasion, with corresponding intervals of deglaciation, we should expect to find, in at least some of the numerous tributary hanging valleys, old drift-filled gorges representing the work of interglacial streams. Such gorges are abundant, though their interglacial origin has not, in most cases, been fully demonstrated. They have been described by Matson[1] as occurring in the Buttermilk valley, and by Tarr, in the valleys of Watkins Glen, Fall Creek, Six Mile Creek, and others in the neighborhood (see references in later paragraphs).

In the autumn of 1909 the senior author discovered in the valley of Six Mile Creek phenomena which apparently had not been recognized before and which seem to have a distinct and important bearing on the problem of multiple glaciation. During the year 1910–11 Mr. Edwin A. Filmer, at the senior author's suggestion, undertook a detailed study of the lower Six Mile Creek valley. The results of this study were embodied in a paper now in manuscript form in the Cornell University Library. The facts brought to light by our investigation of the valley seem worthy of wider publicity than is possible for this manuscript. Hence the present joint paper has been prepared by the senior author, who has, however, made free use of Mr. Filmer's material and conclusions.

[1] G. C. Matson, "A Contribution to the Study of the Interglacial Gorge Problem," *Jour. Geol.*, XII (1904), 133–51.

THE VALLEY OF SIX MILE CREEK AND ITS GORGES

The troughlike Cayuga Lake basin splits at Ithaca into two prominent prongs, the one carrying the Inlet stream from the southwest and the other occupied by Six Mile Creek, which enters from the southeast (see Fig. 1). Of the two, the Inlet valley is the deeper

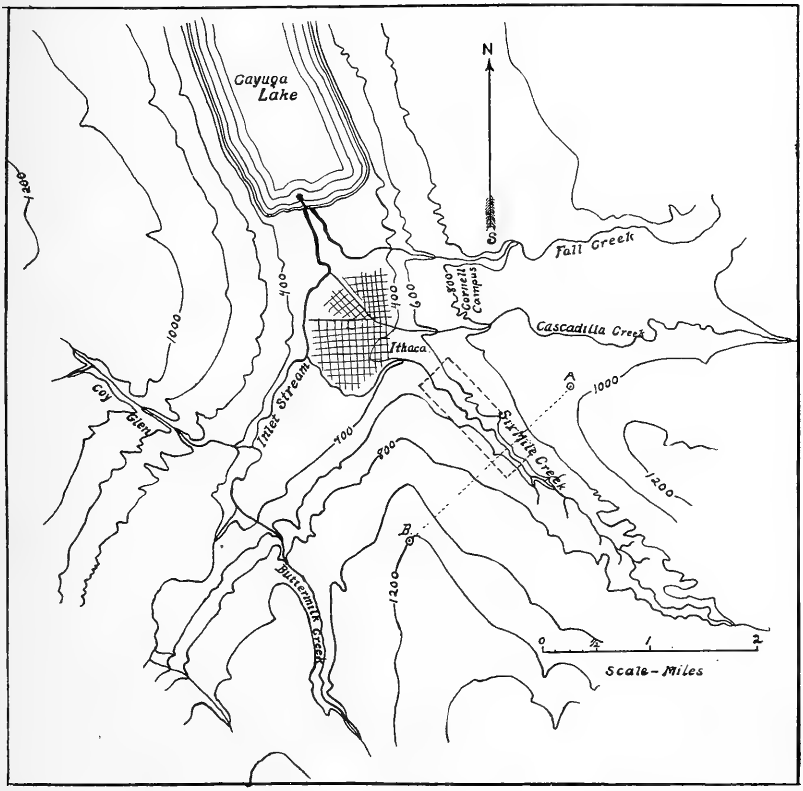

FIG. 1.—Map of the Ithaca region, showing the course of lower Six Mile Creek and its relation to the Cayuga valley and to other streams in the neighborhood. Area shown on Fig. 2 is inclosed by dotted lines.

and the steeper walled and is the main prolongation of the lake basin. It continues with a gradual rise of the valley bottom for about 10 miles southwest of Ithaca and then, by a "through valley" connects southward with one of the tributaries of the Susquehanna. It is a typical glaciated trough. Tributary valleys along this

trough are hanging and their streams enter through narrow rock gorges.

The Six Mile Creek prong is similar to the Inlet trough though it is not so deep and its sides are more flaring. The valley extends southeast from Ithaca in almost a straight line to Caroline Depot, about 7 miles to the southeast, where a southern prong of the valley connects by a low "through valley" past White Church and Willseyville with one of the headwater branches of the Susquehanna leading directly south.

That the valley of Six Mile Creek and this "through valley" were channels of vigorous ice motion during the glacial period is attested by (1) the character of the "through valley" at Willseyville, (2) the trunkated spurs and straight walls of the valley round White Church and Willseyville, (3) the distribution of the moraines in this latter valley and that of Six Mile Creek,[1] and finally by (4) the boat-shaped, typically glaciated cross-section of the lower Six Mile valley, recognized first by Simonds in 1877[2] and later by Tarr.[3] Simonds says (p. 51), referring to the Inlet and Six Mile Creek valleys:

> These deep, well-worn valleys are undoubtedly the result of glacial action. The mass of ice which filled the Cayuga Lake basin divided at its southern extremity. One part, the larger, flowed to the south, wearing down the Inlet valley, and the other traversed the Six Mile Creek valley, both of which were occupied by preglacial streams.

And Tarr (*op. cit.*, p. 20, first paragraph) says:

> . . . . Both Salmon and Six Mile Creek valleys hang at a much lower level than their neighbors (for example Fall, Cascadilla, and Buttermilk [Ten Mile]). I am now convinced that the interpretation of this discordance as opposed to the glacial erosion theory was incorrect and that these two valleys are really confirmatory of the glacial erosion theory. This change in view is the result of a recent study of the valley profiles and a mapping of the morainic deposits of the valleys in question. The latter show that these valleys were occupied by actively moving ice parallel to their axes while the neighboring higher hanging valleys were not. A study of the profiles shows that these discordant hanging valleys have the U-shape of glacial erosion and not the gorge

[1] See Tarr, Watkins Glen-Catatonk Folio.

[2] Simonds, *American Naturalist*, XI (1877), 49–51.

[3] *Jour. Geol.*, XIV, No. 1 (1906), 20.

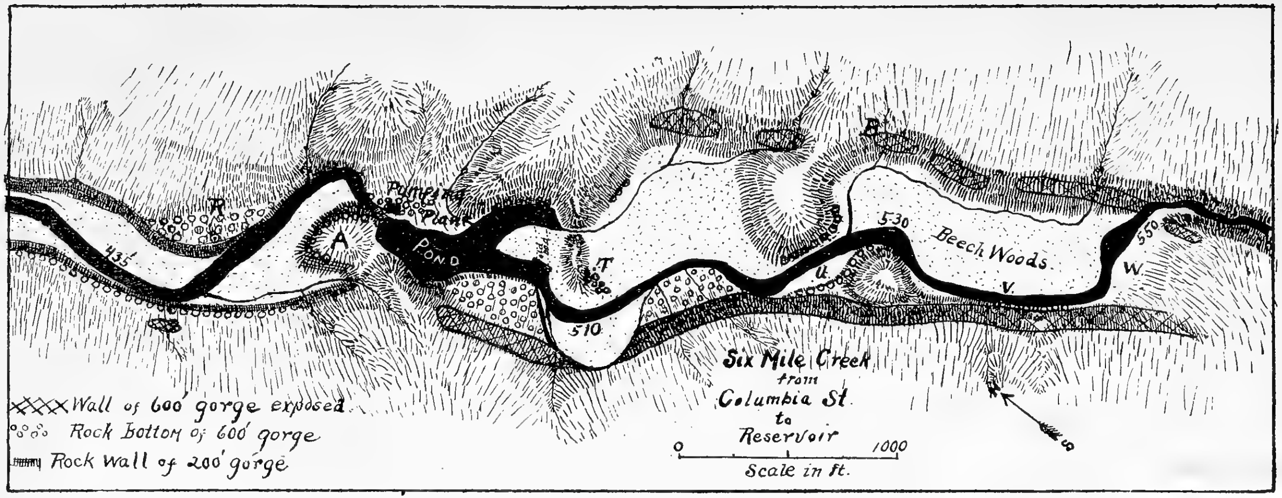

FIG. 2.—Map of a portion of the course of lower Six Mile Creek, showing the topography, the buried gorges, the outcrops of the rock walls and bottoms of the earlier gorges, and the elevations of the stream bed in different parts. The stream flows from right to left of the map. This map is reduced and somewhat modified from a map of the Six Mile Creek valley prepared by the city of Ithaca in connection with surveys for a water supply. Elevations shown represent height above sea-level and were taken from this map. They are approximately correct, but the writers cannot vouch for absolute accuracy.

shape of a rejuvenated valley, the only other explanation which seems a possible one for such discordance. It is believed, therefore, that while the Cayuga valley was profoundly deepened by ice erosion the Salmon and Six Mile Creek valleys were deepened moderately and the Fall, Cascadilla, Buttermilk, and other valleys practically not at all, since they were not occupied by ice freely moving along their axes.

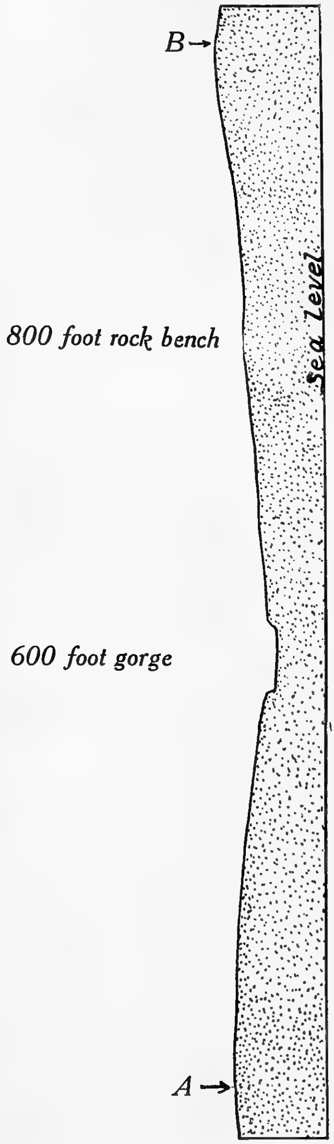

Fig. 3.—A profile across Six Mile Creek valley from A to B of Fig. 1, drawn to natural scale—vertical and horizontal scale the same (2,000 feet to one inch). This profile shows the relation of the 600-foot gorge to the broad valley of Six Mile Creek.

The evidence cited above, together with the foregoing quotations, makes clear the point which we especially desire to emphasize, namely, that the Six Mile Creek valley was occupied by actively moving ice and that it is hanging at a lower level than others near by because glacial erosion deepened its bottom, forming the U-trough so well shown in the profile (Fig. 3).

The rock bottom of the U-trough of lower Six Mile Creek valley is hanging at an elevation about 100 feet above the level of Cayuga Lake, and about 450 feet above the rock bottom of the Inlet valley at this point.

In the bottom of Six Mile Creek trough there is a series of three distinct gorges, which, according to our interpretation, are of different ages. Two are partly, and in places entirely, filled with glacial drift; the third is postglacial. The characteristics of these gorges and their relations to each other

will appear from the following description taken in connection with the map (Fig. 2). There is an older gorge (hereafter referred to from its general width as the 600-foot gorge), 500–600 feet wide, of low gradient, in the bottom of which a second much narrower gorge, which we will call the 200-foot gorge, has been cut to a base-level some 100 feet lower than the other. Both of these gorges antedate the last ice invasion, being still filled by glacial deposits except where they have been cleared out by the postglacial stream.

A third, much younger, gorge (postglacial) is found in places where drift and lake delta deposits have so completely filled the earlier gorges that in re-excavating its valley the stream took a new

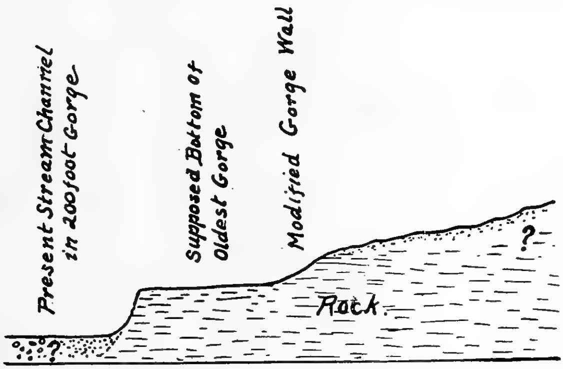

FIG. 4.—Sketch showing bottom and sides of gorge as it appears from upstream

course which, in this case, did not correspond with the former one. In cutting down in this new course the stream, encountering rock, developed the postglacial gorge.

The bottom and walls of the oldest, or 600-foot, gorge are exposed in numerous places for a distance of over a mile. The first evidence is seen just below the Ithaca waterworks pumping plant, where a rock bench, representing the old gorge bottom and east wall, shows in section where the postglacial stream has been undercutting its banks. The characteristics of this section as seen from upstream are indicated in the accompanying sketch (Fig. 4). A little farther upstream, just below the road, is an interesting section best illustrated on the map (Fig. 2). A rock island (*A* of Fig. 2) is found between the buried 200-foot gorge on the south and the recent postglacial gorge. The top of the rock in this island lies approximately

at the level of the previously described rock bench across the stream just below. The rock bench and the rock island together are interpreted as the bottom of the oldest, or 600-foot, gorge. Proceeding upstream about one-quarter mile one comes to a portion of the valley where relations are very clearly shown (see map). Both walls of the old gorge are revealed and the bottom is seen along the southwestern edge and in two places near the middle of the gorge. Proceeding still farther upstream along the northeast bank, one may follow a small stream which enters the amphitheater from the east. For a few hundred feet the course is along a narrow valley excavated in drift. Then, suddenly, one comes upon a nearly perpendicular rock wall which the stream has encountered in its downcutting and over which it falls in a cascade. The rock wall is straight and disappears at both ends under drift. Following southeastward along the projected line of this rock wall, one crosses a hill of drift and soon comes down to the continuation of the wall on the other side (*B*, Fig. 2). Here it is fully exposed, perpendicular, and about 60 feet in height. At the northern end of the exposure the gorge wall is glaciated. The perpendicular cliff is smoothed and striated in true glacial fashion (Fig. 5). From here the gorge wall may be traced almost continuously upstream to the postglacial gorge just below the reservoir, where, after being crossed by the main stream, it disappears under drift (near *W*).

Fig. 5.—The glaciated rock wall of the 600-foot gorge disappearing under drift. From a photograph.

The southwestern wall may be traced with practical continuity all the way. It is, in general, less steep than the northeastern wall.

A steep slope in the place of a vertical cliff is the rule. This southwestern wall also disappears under drift just below the reservoir.

The large amphitheater (just outside the limits of Fig. 2) in which the reservoir lies represents the old gorge again occupied by the stream and locally widened to an unusual extent. Here the gorge walls on the south are steep, but not vertical. On the northeast side the wall appears in several places where small side streams have cut through the overlying drift. It is generally a perpendicular cliff.

The bottom of the old 600-foot gorge, as exposed in a number of places, rises upstream at a rate slightly less than that of the present stream. As a result, the bottom, which appears some 6 feet above the stream at the pumping plant, reaches stream level at the reservoir and lies below it farther up. Partly for this reason the old gorge, in the upper part, becomes more and more indistinct. Its general width, taking the mile of its course just described, is from 500 to 700 feet at the bottom and perhaps 100 feet more than this at the top, for the walls are not everywhere perpendicular. Its relation to the broader glacial trough of Six Mile Creek valley is well shown in the accompanying cross-section (Fig. 3), drawn to a vertical and horizontal scale of 2,000 feet to the inch. The profile makes evident the broad, U-shaped valley with the sharply cut gorge in its bottom. This gorge was probably nearly filled by drift during the various glacial epochs since its formation. In two cases—at the reservoir and south of Green Tree Falls—it is still drift-filled to the top; elsewhere the stream has cleared out a good share of the drift.

In two places on the gorge walls glacial striae have been found; one has already been mentioned, the other is at the eastern edge of the pond, just above the pumping plant. The striations show clearly that the gorge has been covered by ice since its formation, but it seems equally clear from the sharpness of the old gorge walls in most places that the action of this ice was weak and did not materially modify the form of the gorge except at the lower end and where the exposure was particularly favorable. It would seem, therefore, that the gorge must have been cut subsequent to the formation of the broad U-trough of Six Mile Creek valley, in the bottom of which the gorge lies. Otherwise it probably would have

been obliterated by the deep glacial erosion which produced the U-trough.

The second gorge is exposed only below the pumping plant. It is now occupied by the stream all the way from a point a few hundred feet below the pumping plant to the Cayuga Lake delta. Its bottom is nowhere to be seen, since it lies below the level of the present stream. Whether or not it goes below the present base level of Cayuga Lake has not been determined. The fact that the stream is not flowing on a rock bed is not necessarily proof of its having been cut to a lower base level, for it may be that we fail to find rock in the bed of the stream because of the aggradation of the stream bed in connection with the formation of the alluvial fan on which the city of Ithaca stands.

At the Stewart Avenue bridge the gorge is 200 feet wide at the top and has nearly perpendicular walls. Traced upstream, it disappears under drift on the south side of the rock island previously described (*A* of Fig. 2), while the stream lies to the northeast in a postglacial gorge. At the point of disappearance the old gorge has a measured width, from rock to rock of the valley sides, of 125 feet at the bottom. It is nowhere seen farther upstream. It is to be presumed that it lies buried somewhere within the bottom of the older gorge. If so, it must pass under the eastern arm of the pond, for rock exposures preclude its presence elsewhere.

The gorges of the third series are postglacial. They occur where the stream, in cutting down through the later deposits of drift, happened to find itself outside its former gorge. In Six Mile Creek there are three postglacial gorge sections—one just below the pumping plant, another just below the reservoir, and the third just above Green Tree Falls. This latter is being utilized as the site of a second dam for a reservoir for the Ithaca water supply. In the case of the lower of these, at the pumping plant, the diversion of the stream was probably due not so much to filling of the older gorge by glacial till as to the delta deposits built by the stream into Cayuga Lake when it stood at one of its higher levels. A section at the upper end of the rock island by the pumping plant shows till at the base overlain by finely laminated lake clays about 4 feet in thickness, on which lies an undetermined thickness of delta

gravel. This delta deposit is well shown in a cut on the east side of the road, where it was used for gravel. It lies at the general level of the lowest large deltas of Coy Glen and Butternut Creek. There is some evidence, not yet fully worked out, which points to similar deposits as the cause of the diversion of the stream into the two other postglacial gorges farther upstream.

The postglacial gorges are narrow, with walls for the most part vertical rock cliffs. The lower and upper of these gorges still possess waterfalls and cascades. In every case they are distinctly smaller than even the second of the older gorges.

INTERPRETATION

The interpretation put upon the series of gorges in Six Mile Creek is that they were formed by the stream during interglacial intervals. The sequence of events is interpreted as follows:

1. *Preglacial time.*—The stream probably flowed in a broad, mature valley, tributary to the Cayuga Lake trough. The bottom of this valley may have lain at the level of the pronounced rock bench which is clearly marked along the south side of the valley between the 800- and the 1,000-foot contours. This bench is well shown on the topographic map, and is brought out clearly in the accompanying profile (Fig. 3).

2. *First glacial epoch.*—The Cayuga trough, extending parallel with the direction of ice movement, was greatly deepened by ice erosion while at the same time Six Mile Creek valley was deepened considerably and given the flaring U-form typical of glacial erosion. The deepening of the Cayuga trough exceeded that of the Six Mile Creek trough, leaving the latter hanging.

3. *First interglacial interval.*—During this interval Six Mile Creek cut a gorge in the bottom of its hanging valley. That gorge is the oldest, or 600-foot, gorge, described above. The interglacial interval must have been long, for the gorge was cut to an even gradient throughout the whole extent now visible and was also widened very considerably. As we have pointed out before, the gradient of the stream that developed during this interglacial interval was flatter than that of the present stream. An attempt was made, by taking several points as definitely located as possible

on the rock bottom of the old gorge, to plot its profile to scale (see Fig. 6). Points were taken at intervals through a distance of over a mile along the gorge. It was found that, when these points were plotted, the line connecting the first and last passed very close to all the others. This would show that the gorge bottom was well graded. When this profile of the gorge is projected out over the Cayuga valley it is found that at the edge of the Cayuga trough the bottom of the gorge must have been hanging 80 feet above the present lake level. Projecting this line still farther to the axis of the trough, we find it 28 feet above the present lake level. This

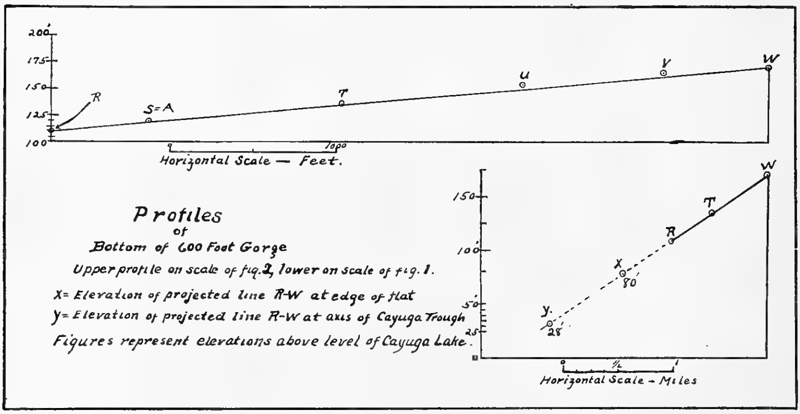

Fig. 6.—Profiles of the rock bottom of the 600-foot gorge. Elevations, taken from the Ithaca water survey map, represent height above Cayuga Lake level. For height above sea add 381 feet.

would indicate that during the first interglacial interval the base level in the main valley stood approximately 80 feet above the present lake level. This base may have been a stream in the main valley or it may have been a lake similar to the one now occupying the basin. We shall discuss later the evidence bearing on this point.

4. *Second glacial epoch.*—A second glacial epoch followed and so lowered the base level in the main valley that the tributary stream again found itself hanging. The ice, during this glacial invasion, seems to have considerably modified the lower part of the 600-foot gorge. The sides and bottom were smoothed and striated and the walls of the gorge at the lower end appear to have been much eroded

and partly destroyed. These gorge walls become practically unrecognizable below the rock island. Without a doubt some of this erosion was done by the later, Wisconsin, ice sheet, but it is thought that the greater part was earlier. The reasons for this belief will be stated in a later paragraph.

5. *Second interglacial interval.*—At the close of the second glacial epoch Six Mile Creek, again left hanging, began cutting another gorge—the 200-foot gorge through which the stream now flows from the rock island to the alluvial flats of the Cayuga valley. The base level of this gorge must have been as low as the present lake level, for rock is nowhere exposed in the stream bed. How much below the present lake level the interglacial base may have been we have no means of knowing. The time interval represented by this gorge must have been much shorter than that during which the older, 600-foot, gorge was cut, for, as has been stated before, the width of the gorge is not much more than one-third that of the older gorge. Judging from the size of the gorge, however, the time must have been much longer than the interval since the last glacial epoch, for the gorge was cut back at an approximately even grade for a distance of two-thirds of a mile at least—how much farther we have no means of knowing, for it is buried under drift. Post-glacial streams of equal size have succeeded, at best, in grading their gorges only a few hundred feet back from the Cayuga valley trough. In fact, most of them come tumbling down over the rocks of the hillsides through gorges which are merely notches in comparison with the interglacial gorges.

6. *The third glacial epoch, Late Wisconsin.*—This ice invasion put an end to the gorge-cutting. It filled or partly filled the older gorge with drift and at its close delta deposits further clogged the gorges until, as the lakes were lowered, the stream found its old channel so completely blocked in places that it must seek lower ground to one side. This determined the stream's course.

The ice of the Wisconsin epoch seems to have done little toward modifying the form of the second gorge. Its walls are still steep and angular and on them we know of no glacial striations having been found. It is the fresh appearance of this second gorge, as compared with the glacially eroded lower end of the 600-foot gorge,

which led us to believe that the Wisconsin ice had played but a small part in the widening and modification of the 600-foot gorge. It seems to be still an open question whether the ice of the *latest* Wisconsin epoch ever extended beyond the belt of strong terminal moraines just south of Ithaca.[1] If it did not, we should expect only feeble erosive action from the Wisconsin ice in the Ithaca region. It is interesting to note, in this connection, that Tarr, in his report on the Watkins Glen-Catatonk Folio, emphasizes the slight erosion by Wisconsin ice.

7. *Postglacial interval.*—This has been marked, in the case of Six Mile Creek valley, by the partial re-excavation of the older gorges and by the cutting of postglacial gorges in places where the postglacial stream found itself outside its former course. The postglacial seems to have been the shortest of the intervals of deglaciation. The gorges are relatively small and show much less stream erosion than the others. In spite of the fact that climatic differences may have influenced the rate of gorge-cutting, it seems legitimate to judge, roughly, the length of an interglacial interval by the size of its gorges.

ALTERNATIVE HYPOTHESES

Before the interpretation outlined above can be considered established, certain questions call for consideration. The first of these is whether the observed relation of the 200-foot gorge to the 600-foot gorge—that is, cut in its bottom—necessitates for its explanation the intervention of a glacial epoch; the second is whether the 600-foot gorge may not be a product of preglacial rejuvenation.

In answer to the first of these questions it may be said that the whole problem resolves itself into a search for an adequate reason why the stream, after cutting the 600-foot gorge to a low gradient—evidently to its local base level—and widening it considerably, should suddenly have renewed its downcutting sufficiently to have formed the 200-foot gorge in the bottom of the older one. Several possible explanations present themselves: (*a*) there may have been a lake in the main valley whose level was suddenly lowered; (*b*) the

[1] Chamberlin, in his report on the "Terminal Moraines of the Second Glacial Epoch" in the *Third Annual Report of the U.S. Geol. Survey,* states, as a possibility, that the moraine in question represents the limit of Wisconsin ice movement. It seems to the writer that this possibility has never been satisfactorily set aside.

main valley may have been occupied by a stream which in some way, perhaps by uplift, was suddenly rejuvenated; (*c*) a glacial invasion may have altered the base level in the main valley. In regard to these possible explanations, we believe it not only possible, but very probable that a lake, similar to the present Cayuga, existed in the valley previous to the last ice invasion. It is conceivable, also, that such a lake might be drained comparatively suddenly. For instance, a hard rock stratum might be so disposed as to form a barrier which would yield rather suddenly and expose weaker rocks beneath. It is evident, therefore, that we cannot at once dispose of hypothesis (*a*). On the other hand, we have in Six Mile Creek alone no means of proving that a river in the main valley may not have been suddenly rejuvenated. It will be apparent, however, that if such a trunk river, by its rejuvenation, led to the formation of a gorge in Six Mile Creek valley, it should, in a similar manner, have led to the formaton of similar gorges, corresponding to the 200-foot gorge of Six Mile Creek, in all the other tributaries. We should expect to find, then, that the wide buried gorge of Buttermilk Creek, which apparently corresponds to our 600-foot gorge, should have a narrower gorge, corresponding to our 200-foot gorge, sunk in its bottom. Such is not the case, however. We must conclude from this, we believe, that hypotheses (*a*) and (*b*) are both untenable, for the argument stated above would apply equally well in the case of a lake whose level was suddenly lowered. The third hypothesis suggested—the lowering of the base level of the tributaries by the intervention of a period of glacial erosion—does not meet with this objection, for it might well happen that the Six Mile stream, after the disappearance of the ice, would find itself still in its old channel, while the Buttermilk stream, happening to find its old channel blocked by drift or delta deposits, cut a new gorge elsewhere, as, in fact, it has done.

A careful study of the buried gorges of Butternut Creek, Newfield Creek, and others should be made in order to settle the question beyond possibility of dispute.[1]

[1] In this connection it may not be out of place to call the attention of any whose lot it may be to make a further study of these buried gorges to a remarkably fine buried

To return to a consideration of the second question, namely, whether the 600-foot gorge may not be a product of preglacial rejuvenation, we have seen that lower Six Mile Creek valley was profoundly deepened and modified by ice erosion and that it was given the broad, flaring U-form of a typical glacial trough. The 600-foot gorge now lies in the bottom of this trough and still retains perpendicular or nearly perpendicular rock walls throughout much of its extent. If this gorge was formed by a preglacial stream it must have survived the deep scouring which lowered the bottom of the valley to an extent of some 200 feet or more and modified its form to that of a typical U-trough. It is hard to believe that such a gorge would not have been widened by the ice and made a part of the U-trough had it been in existence while the erosion was in progress. At least the sharpness of the gorge walls would have been destroyed. It is true that, on the southwest wall of the gorge, this sharpness has been to some extent destroyed, but this is probably the result of erosion by the ice of some of the later periods. The northeast wall is still sharp.

A further point which may have a bearing on this problem is the fact that in the valley of Fall Creek there are drift-filled gorges, as, for instance, those just north of the Triphammer Bridge, whose bottoms are nearly 200 feet above that of the 600-foot gorge in the Six Mile Creek valley. Certainly these gorges cannot belong to the same cycle as the 600-foot gorge, for both seem well graded, yet their base levels are so different. If either is preglacial it must be the higher one. Buried gorges at levels higher than our 600-foot gorge are to be found in several of the valleys of the region besides that of Fall Creek. On detailed study, with careful leveling, it may appear that even the 600-foot gorge of Six Mile Creek belongs to a

gorge in the valley of a small stream which enters Cayuga Lake from the east at Shurger Point. In following up the gorge of this stream one soon encounters a series of cascades and falls aggregating perhaps 75 feet in height. Above these cascades is a gorge 100 feet, more or less, in width at the bottom, with rock walls, and a remarkably even, low-gradient rock floor. The gorge, with these characteristics, continues upstream for half a mile to a point just below the trolley track (one mile from the lake), where it suddenly ends, the rock walls disappearing under drift, and the stream tumbling in over a fine fall 60–80 feet in height. The bottom of the upper section of this gorge has a low gradient and hangs at a level somewhere between 60 and 100 feet above that of Cayuga Lake.

comparatively late interglacial interval, and that some of these higher gorges represent still earlier intervals, thus still further increasing the complexity of the glacial record.

LITERATURE

The interpretation of these gorges which we have presented is not, in all respects, new. A brief summary of the literature pertaining to Six Mile Creek and its gorges will indicate the status of the problem previous to the recent investigations.

Probably the first mention of the drift-filled gorges was made by Simonds in 1877,[1] who says: "The valley of Six Mile Creek furnishes some special examples of the drift phenomena. In several places its old channel has been completely choked up with masses of morainic débris, about which the present stream has been obliged to cut its way through deep canyons." This reference is interesting, though it contains nothing bearing directly on the problem in hand.

In the *Physical Geography of New York State* (New York: Macmillan Co.), published in 1902, Tarr, under the heading "Interglacial(?) Gorges" (pp. 178–79), calls attention to "numerous gorges which are broader than the postglacial valleys and partially obscured by glacial till, showing that they were formed either during preglacial or interglacial times," and mentions Six Mile Creek as an especially good example. The problem which has ever since been recurring, and, in fact, still calls for discussion, is there clearly stated: "Were the gorges [of central and western New York] due to the interglacial conditions or to an uplift in preglacial times?"

In a paper on "Hanging Valleys in the Finger Lake Region of Central New York"[2] Tarr discusses the Six Mile Creek gorges at some length. He points out the fact that practically all the valleys tributary to the Finger Lakes possess gorges "which antedate the last advance of the ice" and that in the case of Six Mile Creek

> the stream alternately enters the buried gorge, forming broad amphitheaters, and where it for a short time leaves the earlier gorge, crosses spurs of rock in

[1] *American Naturalist*, XI (1877), 49–51.

[2] *American Geologist*, XXXIII (May, 1904), 271–91.

narrow, postglacial gorges. . . . . A third condition is where a stream enters the earlier gorge, clears out part or all of its drift filling, and follows it across the steepened slope to the main valley. This is certainly the case in lower Six Mile Creek.

It should be noted that in this reference to Six Mile Creek Tarr makes no distinction between the gorge characterized by the "broad amphitheaters" mentioned above (our 600-foot gorge) and the gorge of "lower Six Mile Creek," which the stream follows to Cayuga valley (our second, or 200-foot, gorge).

A following paragraph (p. 281) is especially significant in connection with the facts brought out by our recent study. We quote in full:

This evidence establishes a third point regarding the drainage history of Seneca and Cayuga valleys, namely, that there is an almost, if not absolutely, uniform condition of gorges in the bottoms of the mature hanging tributary valleys; that these gorges, being drift-filled, antedate the last ice advance; that they are broader and deeper, hence required longer time to form than the postglacial gorges; and that, where they enter the main valleys, their rock bottoms are above lake level. They therefore resemble hanging valleys, since their bottoms are in some cases, as Taughannock, about 400 feet above the main valley bottom one-half mile distant. Since we have no data proving what their bottom slope actually is, though it seems evident that their slope is very steep, and possibly great enough to carry them down to the main valley axis, it may not be proper to consider them hanging valleys. If the interpretation of hanging valleys is warranted, which I doubt, then the tributary valleys to Seneca and Cayuga Lakes are double hanging valleys—an upper mature hanging valley and a lower hanging gorge valley.

In the case of Six Mile Creek at least we now have the data proving the actual bottom slope (Fig. 6) and it appears that the slope is, in reality, comparatively gentle, and that the double hanging valleys of which Tarr speaks are, in fact, present.

In another paragraph in the same paper (p. 284) the results of our recent study are foreshadowed in the following words:

A modification of the glacial erosion theory has been advanced during the progress of the investigation of the problem and is still being considered. It is as follows: During its first advance the ice deeply eroded the valleys; during interglacial conditions the older valleys were cut; with return of glaciation the valleys were deepened still further. During as many glaciations as this region experienced this process was continued. On this basis the older gorges are interglacial; their cause is the lowering of their base level by the overdeepening of the valleys to which they were tributary. Since facts sufficient to establish

or to overthrow this explanation are not yet at hand it must stand at present merely as a working hypothesis.

The question of the explanation of the buried gorges by regional rejuvenation is then raised (p. 290). Two sentences in connection with this are quoted:

Distinct progress toward solution might be made if it were possible to establish the grades of the buried gorges. If they can be shown to have such steep grades as to carry them down to the main valley bottoms the theory of rejuvenation will be greatly strengthened.

In this connection we again call attention to Fig. 6, showing the projected grade of the rock bottom of the oldest of the buried gorges of Six Mile Creek, which profile, it would seem, does not tend to strengthen the rejuvenation theory.

As late as 1906 we find the following in regard to the gorges of the Finger Lake valleys:[1]

There are some facts which indicate possible greater complexity of ice erosion, for in some of the valleys there is apparently more than one buried gorge; but the evidence on this point is not as yet convincing, and for the present we can point with certainty to no greater complexity than that of two periods, one the Wisconsin, the other of some one of the earlier ice advances, with which the work of the glacial geologists of the Mississippi Valley have made us familiar.

Tarr's latest published conclusions in regard to these buried gorges may be found in the Watkins Glen–Catatonk Folio of the United States Geological Survey, 1909. He concludes that evidence of a pre-Wisconsin period of ice erosion "is afforded by the presence of hanging gorges, partly buried in Wisconsin deposits. . . . . These are evidently interglacial gorges cut in the bottoms of hanging valleys that were left hanging by overdeepening of the main troughs through ice scouring" (p. 117, field edition); and again, p. 224: "We have conclusive evidence here [referring to the buried gorges] of only two advances, but there may have been more."

Matson (*op. cit.*) at an earlier date had described buried gorges in the valley of Buttermilk Creek, near Ithaca, which led him to the conclusion that in that valley there exists a "series of complex gorges which are considered interglacial. The minimum number of epochs of deglaciation is two; the maximum number, four. . . . ."

[1] R. S. Tarr, "Watkins Glen and Other Gorges of the Finger Lake Region of Central New York," *Pop. Sci. Mo.* (May, 1906), 387–97.

It will be evident from the foregoing summary of the literature that more than one set of buried gorges has been recognized, or at least suspected. The problems still left unsolved were: (1) the details of these gorges, and the number present in different valleys; (2) the gradient of the gorge bottoms; (3) the possibility of the explanation of one of the buried gorges by preglacial rejuvenation.

BROADER BEARINGS OF THE PROBLEM

The buried gorge phenomena of Six Mile Creek do not stand isolated. Similar phenomena in many other valleys of the region have been recorded.

If our interpretation is correct they all form a definite part of the sequence of events connected with the glacial epochs. Some valleys should show several gorges; some, perhaps, in cases where the stream happened to find the same channel each time, should show only one. The phenomena in Six Mile Creek valley point to at least three glacial epochs separated by periods of time longer than that since the last epoch. Other valleys may show an even more complete record. Buttermilk valley shows a more complex series than Six Mile, and Butternut Creek has a splendid series of old gorges, very little studied as yet.

As the study of these gorges progresses it should be possible to settle many points now in question. For instance: the 600-foot gorge of Six Mile Creek was cut to a base level approximately 80 feet above present lake level. The largest gorge of Buttermilk was cut to approximately the same base. The base of the largest gorge in Butternut Creek lies at approximately the same level—perhaps somewhat higher. By careful leveling it should be possible to determine with close approximation the base level to which each of these gorges was cut. It may be possible in this way to determine whether a lake (horizontal base level) or a river (inclined base level) occupied the main valley at the time of the cutting of the gorges. The Finger Lake valleys and their tributaries afford abundant evidence which must be studied carefully before a final decision as to the full complexity of the glacial period in central New York can be made.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA FOR 1909, 1910, 1911, AND PART OF 1912

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CONTENTS

I. GENERAL

1909

Adams[1] maintains that the pre-Cambrian may be divided on the basis of major diastrophic movements into Eo, Meso, and Neo-Proterozoic, and that this basis may serve the purpose of international correlation better than comparative study of unconformities. He objects to the two fold classification of the pre-Cambrian suggested by Van Hise, on the ground that the break at the base of the Animikie is as great as that which follows the Keewatin.

Adams[2] states that amphibolites consisting of aggregates of hornblende, pyroxene, mica, and other minerals, in various proportions, have developed from (*a*) the metamorphism and recrystallization of impure calcareous sediments; (*b*) alteration of certain basic dikes; and (*c*) alteration of limestones by batholithic intrusions of granite. The granites show marginal digestion of limestones

[1] F. D. Adams, "The Basis of pre-Cambrian Correlation," *Jour. Geol.*, XVII, No. 2 (1909), 105–23.

[2] F. D. Adams, "Origin of Amphibolites of the Laurentian Area of Canada," *Jour, Geol.*, XVII, No. 1 (1909), 1–18.

grading into a mosaic of hornblende and feldspar crystals. With diminishing hornblende the gradation passes into a pyroxene, scapolite, biotite rock, and finally into coarse crystalline limestone.

From analyses of limestones representing three stages of alteration to amphibolite near a contact, Adams concludes that the alteration has involved the introduction from the granite of silica, alumina, iron, magnesia, and alkalies, and the loss of lime and carbonic acid. This conclusion seems to rest on the assumption that the alteration took place without change of mass.

Lindgren[1] presents a résumé on the principal epochs of the segregation of metals on the North American continent. Many different epochs of ore formation are embraced in the pre-Cambrian period (in many different places). In the eastern part of the continent iron, copper, nickel, silver, and gold ores occur in association with a variety of both intrusive and extrusive igneous rocks ranging from granite to basalt. In the Cordilleran region, intrusives, principally granites, are almost exclusively represented. Intrusive diorite, gabbro, and diabase are present. The Cordilleran pre-Cambrian ores are principally gold and silver. The lead and zinc ores of the pre-Cambrian are unimportant.

Van Hise[2] maintains that the bases of classification of the pre-Cambrian should be physical, those of most importance being: (1) lithologic character, (2) continuity of formations, (3) likeness of formations, (4) like sequence of formations, (5) subaerial or subaqueous deposits, (6) unconformities, (7) relations to series of known age, (8) relations with intrusive rocks, (9) amount of deformation, (10) degree of metamorphism. His major divisions of the pre-Cambrian are Archaean and Algonkian.

1910

Adams[3] concludes that large igneous intrusives rising from the deeper portions of the earth are one of the principal agents of the

[1] Waldemar Lindgren, "Metallogenetic Epochs," *Econ. Geol.*, IV, No. 5 (1909), 409–420.

[2] C. R. Van Hise, "Principles of Classification and Correlation of the pre-Cambrian Rocks," *Jour. Geol.*, XVII, No. 2 (1909), 97–104.

[3] Frank D. Adams, "The Origin of the Deep-seated Metamorphism of the pre-Cambrian Crystalline Schists," *Compte Rendu*, Congrès Géologique International, 1910, pp. 563–72.

metamorphism of rocks, since more and more cases of intimate relationship between the occurrence of schists and batholithic intrusions, principally granite, have been found as geologic study proceeds.

Coleman[1] presents a summary of the stratigraphy and correlation of the pre-Cambrian as adopted by the international committees on correlation, before the British Association at Winnipeg.

Coleman[2] argues that the abundance of carbonaceous shales, limestones, and other ordinary sediments in the Keewatin of Ontario implies the existence of life and climatic conditions not unlike the present.

Coleman[3] infers that uniformitarian conditions prevailed in the pre-Cambrian from the similarity of pre-Cambrian sediments to those of later periods.

Coleman[4] presents a summary on the degree and kind of metamorphism of the various pre-Cambrian formations of northern Ontario. The agents of metamorphism as outlined by Coleman are heat, pressure, and solutions.

Coleman[5] presents a historic summary of the methods which have been employed in classifying the Archaean of Ontario. The principal methods, according to Coleman, in the order of their historic development are relative stratigraphic position of formations, lithologic similarity, unconformity, eruptive contact, and basal conglomerates.

Coleman[6] summarizes the principal known events in the history of the Canadian shield.

[1] A. P. Coleman, "The pre-Cambrian Rocks of Canada," *British Assoc. Adv. Sci. Rept.* 79th Meeting, 1910, pp. 474–75.

[2] A. P. Coleman, "Climate and Physical Conditions of the Keewatin," *Geol. Soc. Am. Bull.*, XXI, No. 4 (1910), 778–79.

[3] A. P. Coleman, "The Bearing of pre-Cambrian Geology on Uniformitarianism," *British Assoc. Adv. Sci. Rept.*, 79th Meeting, 1910, pp. 473–74.

[4] A. P. Coleman, "Metamorphism in the pre-Cambrian of Northern Ontario," *Compte Rendu,* Congrès Géologique International, 1910, pp. 607–16.

[5] A. P. Coleman, "Methods of Classification of the Archaean of Ontario." *Compte Rendu,* Congrès Géologique International, 1910, Vol. I, pp. 721–33.

[6] A. P. Coleman, "The History of the Canadian Shield," *Nature,* LXXXIV (September 15, 1910), 333–39.

R. A. Daly[1] believes that the pre-Cambrian ocean probably contained less lime than the present ocean, for the following reasons: the sea must have been fresh or nearly fresh initially; the pre-Cambrian rivers probably carried less lime because the lands of that time did not possess the great development of limestone of the present; the lime contributed to the sea was probably precipitated by ammonium carbonate generated through the decay of organisms which accumulated on the sea bottom because of the lack of scavengers. The field evidence which he cites in favor of this hypothesis is the fineness of grain of many undisturbed pre-Cambrian dolomites which he believes is indicative of chemical precipitation, and the general scarcity of fossils in pre-Cambrian rocks, which he ascribes to the lack of lime in the sea from which organisms could obtain the materials for shells and hard parts.

Evans[2] believes that the sudden appearance of the Cambrian fauna may have been due to causes very similar to those which gave rise to the sudden appearance of the Tertiary fauna. These periods were preceded by great uplift of the lands, vulcanism, and widespread aridity. The record of the evolution of life immediately preceding them is therefore largely lost or inaccessible.

Harder[3] states that the pre-Cambrian manganese ores of the United States are found in association with the hematite deposits of the Lake Superior region; as manganiferous zinc ores at Franklin Furnace, N.J.; and in the metamorphic rocks of the Piedmont region of Virginia, North Carolina, South Carolina, and Georgia.

Hayes[4] states that the pre-Cambrian iron ores of the Lake Superior region furnish about 80 per cent of the annual ore production of the United States.

The iron formations from which the ores were derived are sedimentary deposits, consisting mainly of cherty iron carbonates,

[1] R. A. Daly, "Some Chemical Conditions in the pre-Cambrian Ocean," *Compte Rendu*, XI. Congrès Géologique International, 1910, pp. 503–9.

[2] John W. Evans, "The Sudden Appearance of the Cambrian Fauna," *Compte Rendu*, XI. Congrès Géologique International, pp. 543–46.

[3] Edmund Cecil Harder, "Manganese Deposits of the United States with Sections on Foreign Deposits, Chemistry, and Uses," *Bull. 427, U.S. Geological Survey*, 1910, pp. 298.

[4] C. W. Hayes, "Iron Ores of the United States," *Bull. 394, U.S. Geological Survey*, 1910, pp. 76–84.

ferruginous cherts, and ferrous silicate rocks interbedded with clastics and extrusives. Their average iron content ranges from about 25 to 35 per cent iron. The ores were developed by the solution and removal of silica, and the oxidation, solution, transportation, and redeposition of the iron, in the exposed portions of the formations, having structural peculiarities favorable to the circulation of water, such as jointing, pitching basements, faults, etc.

The ores mined are mainly hematite and limonite, although some magnetite has been developed through deep-seated conditions of high pressure and temperature. Their iron content, while steadily falling, has generally been above 50 per cent, but 40 per cent ores are now mined as mixers for higher-grade materials. The estimated tonnage of the available ores, including all ores above 55 per cent iron, and 25 per cent of all formations containing from 45 to 55 per cent iron is 3,500,000,000 tons. Iron formations containing from 35 to 45 per cent iron are classed as available in the future. Their tonnage is enormous, 72,000,000,000 tons being taken as an arbitrary estimate.

The pre-Cambrian Adirondack area of New York contains commercial deposits of non-titaniferous magnetites, titaniferous magnetites, and red hematites.

The non-titaniferous magnetites are intimately associated with sedimentary rocks, marble, schists, and gneisses. Separation is generally made by magnetic methods and the ore shipped to the iron centers, excepting a small tonnage which is smelted in local coke ovens. The ores are both Bessemer and non-Bessemer. The lowest grades contain about 35 per cent iron when mined, the richest from 60 to 65 per cent. The concentrates run from 60 to 65 per cent. No fair estimate of the tonnage can be made. The known deposits of high grade are believed to contain 35,000,000 tons, and the leaner deposits carrying over 35 per cent iron may yield 75,000,000 tons of concentrates.

The ores are believed to have been developed in connection with the intrusion of igneous rocks.

The titaniferous magnetites carry at least 8 per cent of titanic acid and on an average about 15 per cent. They are found at the margins of a gabbro-anorthosite area, and are spoken of as magmatic

segregations. The richest ores contain about 60 per cent iron. They are not used at present because of their titanium content, excepting on a very small scale. Their extent is unknown. In the Lake Sanford district alone, 90,000,000 tons are known.

Karl L. Henning[1] reviewed the principal results of Van Hise and Leith's *Pre-Cambrian Geology of North America.*

Kemp[2] states that the following tentative chronologic table of the Swedish pre-Cambrian based on the work of Sederholm and Högbom is finding favor:

Upper or Jotnian	Epijotnian dislocations Jotnian Subjotnian land surface denudation and igneous rocks
Middle or Jatulian	Epijatulian folding Jatulian Subjatulian land surface and denudation
Lower or Archaean	Serarchaean granites Archaean No chronological subdivision. Difference due to varying metamorphism

Kemp shows that the Archaean of Sweden is separable into two major divisions, an older complex of minor sediments and major deep-seated intrusives, and a younger series of intrusive granites consisting of four distinct types. Similarly, the Archaean of the Lake Superior region embraces an older division consisting of minor sediments and major igneous rocks, mostly basic extrusives, however, and a younger division consisting almost entirely of granitic intrusives. The sediments of the Swedish Archaean contain squeezed conglomerates, crystalline limestones and dolomites, mica schists, leptites, garnet, cordierite, sillimanite, and graphite gneisses of probable sedimentary origin. The leptites are fine-grained, banded, and stratified rocks, which include sediments and iron ores. The Archaean sediments of the older division in the Lake Superior region include iron formation and minor clastics. The Archaean igneous rocks of Sweden of the older division comprise granites, gneisses, and gneissoid intrusives, titaniferous ore-bearing gabbros,

[1] Karl L. Henning, "Die pro-kambrische Geologie von Nord Amerika," *Naturwissenschaftliche Wochenschrift*, N.F., Bd. No. 28 (July, 1910), pp. 433–37.

[2] J. F. Kemp, "Comparative Sketch of the pre-Cambrian Geology of Sweden and New York," *New York State Museum Bull. 149*, 1910, pp. 93–119.

syenites, and anorthosites. The leptites, sediments, and iron ores are compared by Kemp to channels amid an archipelago of large island intrusives.

The surface on which the Jatulian was deposited is regarded as hilly by Högbom and mountainous by Törnebohm. Kemp believes that the time gap of erosion which separated the Jatulian from the Archaean was not so long as the one which separated the Cambrian from the Jotnian or the Jotnian from the Jatulian. The Jatulian in Sweden consists of quartzite, schists, dolomitic limestones, and beds of anthracite, which have a maximum thickness of two meters.

Professor Sederholm has divided the Jatulian of Finland into an upper and lower member which consist of eruptives, greenstones, clastics, and dolomites, and has compared it to the Upper Huronian of the Lake Superior region. The Jatulian closed with a period of very intense folding, followed by intrusions of a peculiar porphyritic granite which decomposes readily, forming the so-called *rapakivi*, Finnish for rotten stone. This has afforded an important guide by which Swedish geologists have measured the age of various other intrusives and metamorphic rocks. The erosion which followed developed an even floor free from weathering products, on which the Jotnian diabases and ripple-marked and sun-cracked sandstones were deposited. These now lie in isolated patches. They are compared by Professor Kemp to the Torridonian of Scotland and the Keweenawan of the Lake Superior region. Finally, Cambro-Silurian sediments of which widespread relicts remain were deposited upon a generally even surface covered with a weathering breccia which grades downward into a kaolinized gneiss. So-called sandstone dikes containing Cambrian fossils are found in gneisses far from Cambrian beds.

Most of the comparisons made by Professor Kemp between the pre-Cambric of Sweden and New York are necessarily petrographic and appeal to those who are personally familiar with the rocks of the New York pre-Cambrian.

Lane[1] finds that the grain of Laurentian granites tends to be uniform from the margin to the center, from which he infers that

[1] A. C. Lane, "The Stratigraphic Value of the Laurentian," *Compte Rendu*, XI. Congrès Géologique International, 1910, pp. 633–37.

the country rock which the granites invaded must have been nearer the condition of crystallization or fusion than the granites, and that the Laurentian granites are batholiths rather than re-fused sediments or portions of the original crust of the earth.

Lindgren[1] states that the principal pre-Cambrian ores of North America are iron, copper, nickel, gold, and silver. The iron ore types are: magnetite and ilmenite of igneous origin, and hematites of sedimentary and igneous origin, which have been concentrated by surface waters. The copper, nickel, and silver of the Lake Superior region and the copper ores of the South are connected with basic igneous rock. The gold quartz veins of the south are related to granitic intrusions.

Newland[2] states that the iron ores mined in Sweden are all pre-Cambrian magnetites associated with crystalline schists and acid igneous rocks. There are also some low-grade, gabbroic, titaniferous magnetite silicate rocks, and bog lake deposits, the latter famous as exemplifications of present-day ore deposition, but these are not mined.

Most of the ore comes from central Sweden. A considerable portion of this is unique for its low phosphorus content, and because of its exceptional qualities can be mined from very small deposits on a miniature scale. The ore is reduced in charcoal furnaces. Nearly half of the output of central Sweden is high phosphorus ore from Grangesberg. The high phosphorus ores are all exported, since there are no coals in Sweden suitable for furnaces. The ores of central Sweden occur in association with sediments, leptites, and gneisses which form winding lenses between massive intrusives. The ores at Norberg and Striberg consist of banded magnetite quartz rocks with a sedimentary aspect.

The Kiruna and Gellivare magnetite apatite deposits of Lapland resemble the titaniferous ores of New York in their association with sodic rocks. The ores at Gellivare are lenses, bands, and chimneys in syenite, the entire mass showing regional metamorphism. The

[1] W. Lindgren, "Metallogenetic Epochs," *Jour. Can. Min. Inst.*, XII (1910), 102–13.

[2] D. H. Newland, "Notes on the Geology of the Swedish Magnetites," *New York State Museum Bull. 149*, 1910, pp. 107–19.

ores at Kiruna are of the nature of large dikes between igneous rocks. The largest, that of Kurunavaara, is 5 kilometers long and from 50 to 164 meters thick, and lies between a quartz porphyry hanging wall and syenite porphyry foot. The ore has a dense, steely appearance, and consists of magnetite with a little apatite. They are probably igneous in origin because of the igneous character of the wall rocks, the igneous texture of the ore, the presence of apatite dikes in the ore and wall rock, and the existence of apophyses of ore in the footwall and the surrounding porphyry.

Sederholm[1] believes that certain pre-Cambrian graphitic schists and limestones indicate the existence of progonozoique forms. The progonozoique formations alluded to are the formations which contain vestiges of the ancestral forms of the life which suddenly appears in the Cambrian.

Sollas[2] ascribes the scarcity of fossils in the pre-Cambrian to a feeble development of hard parts in the organisms of that period. From the rough approximation that "the ontogeny of the individual repeats the phylogeny of the race," he infers that the organisms of the pre-Cambrian were similar to the larval forms of the existing invertebrata. Such larval forms are not known to have been preserved, notwithstanding that some are endowed with calcareous skeletons.

Walcott[3] accounts for the sudden appearance of abundant marine life with the Cambrian by assuming that marine life first developed in the open ocean and did not migrate to the shore zones until the Algonkian uplift preceding the Cambrian had largely excluded the seas from the continents.

Walther[4] ascribes the scarcity of fossils in certain pre-Cambrian formations immediately below the Cambrian to their deposition

[1] J. J. Sederholm, "Sur les vestiges de la vie dans les formations progonozoique," *Compte Rendu*, XI. Congrès Géologique International, 1910, pp. 515–24.

[2] W. J. Sollas, "The Fauna of the Protaeon," *Compte Rendu*, XI. Congrès Géologique International, 1910, Vol. I, pp. 499–501.

[3] C. D. Walcott, "Abrupt Appearance of the Cambrian Fauna on the North American Continent," *Smithsonian Misc. Coll.*, LVII, No. 1 (1910), pp. 1–16, 1 pl., 1 fig.

[4] T. Walther, "Die lithologischen Eigenschaften der Gesteine im Liegenden der kambrischen Formations," *Compte Rendu*, XI. Congrès Géologique International, 1910, Vol. I, pp. 511–513.

under desert conditions. He applies this hypothesis to the Algonkian of the Grand Canyon, the Torridonian sandstone of Scotland, and the Sparagmite formation of northern Scandinavia.

Matthew[1] ascribes the sudden appearance of Cambrian life to the obliteration of fossils by chemical changes and other causes, partly to the absence or rarity of Benthos in the earlier faunas, and partly to the fact that fossiliferous strata beneath the Cambrian are arbitrarily attached to the base of the Cambrian.

1911

C. K. Leith and E. C. Harder[2] state that Brazilian hematite ores of Minas Geraes are found about 300 miles from the coast, in a region constituted by a basement complex of crystalline schists, upon which rests a sedimentary series consisting of slates, schists, quartzites, ferruginous quartzites or itabirites with interbedded hematite ores, gneiss, ferruginous carbonate rocks, carbonate rocks, and amphibolites. The sediments have been severely folded and altered by granitic intrusives. Near the bottom quartzite predominates. Argillaceous sediments predominate near the top and the iron ore beds are most abundant near the middle of the sedimentary series.

The most abundant ores of the district are thin-bedded, fine-grained hematites, of excellent grade. Thick-bedded, massive, nearly pure hematites rank next in importance.

A very large tonnage of fragmental ore consisting of residual hematite and more or less transported débris resulting from the breakdown of the ore deposits is in sight. These ores are cemented with sands, clays, and limonite and therefore are lower in grade than bedded deposits. A negligible quantity of ore is known to have developed from the leaching of ferruginous carbonates and itabirite.

The Lake Superior ores have resulted from primary deposition and secondary concentration. In no known case has primary deposition alone developed ore. All the ores are related to structu-

[1] G. F. Matthew, "The Sudden Appearance of the Cambrian Fauna," *Compte Rendu*, Congrès Géologique International, 1910, pp. 547–551,

[2] C. K. Leith and E. C. Harder, "Hematite Ores of Brazil and a Comparison with Hematite Ores of Lake Superior," *Econ. Geol.*, VI (1911), 670–86.

ral features which favored the circulation of water. In Brazil, the iron ores are the result of primary deposition, and while some secondary concentration by leaching has taken place, it is local and very subordinate. The fragmental ores of Brazil have no counterpart in the Lake Superior region excepting in conglomerates at the base of formations resting on ore deposits. Glacial erosion may have removed large quantities of fragmental ore in the Lake Superior region.

Miller[1] points out that certain pre-Cambrian stratigraphic units of the Northwest Highlands of Scotland are similar to some of Canada as indicated by the following table:

Canada	Scotland
A. Keweenawan.	A. Torridonian.
B. Huronian with intrusives.	B. Intrusives of Lewisan. A fragmental series of quartzite, etc., has been removed before Torridonian was deposited.
C. Keewatin-Laurentian Complex. Granite and gneiss, greenstone, limestone, iron formation, etc.	C. Lewisan. Fundamental Complex. Gneiss, hornblende and chloritic schists, limestone, and iron formation.

Miller and Knight[2] believe that the present use of the term Laurentian as applied to the granites and granite gneisses intrusive into the Keewatin but presumably not in the Lower Huronian may have led to poorly based correlations.

[1] W. G. Miller, "A Geological Trip in Scotland," *Ont. Bur. Mines, 20th Ann. Rept.*, 1911, pp. 259–69.

[2] W. G. Miller and C. W. Knight, "The Laurentian System," *Ont. Bur. of Mines, 20th Ann. Rept.*, 1911, pp. 280–84.

[*To be continued*]

Reviews

West Virginia Geological Survey. Marion, Monongalia and Taylor Counties. By RAY V. HENNEN. 1913. Pp. 844, pls. 33, figs. 11, maps 3; Cabell, Wayne and Lincoln Counties. By C. E. KREBS. Pp. 483, pls. 26, figs. 6, maps 9.

The report on the counties of Marion, Monongalia, and Taylor discusses the industrial development, the physiography, the geology, and the mineral resources of this district. These counties are in the northern part of the state in the heart of the Appalachian coal field. They are exceedingly productive of gas and oil and are rich in building stones, glass sands, and clays. To all this must be added great agricultural fertility.

The surface of the region has been eroded to a strikingly mature stage; it is almost entirely a country of steep slopes and very narrow divides. Structurally this district forms part of the eastern flank of the Appalachian Basin. The formations have a general western dip which is modified by local folds. The structure contours on the geological map make it evident that the gas has generally collected in the tops of the anticlines, and that the oil has been found most abundantly upon the flanks of the anticlines. A general section of the strata is as follows:

Upper Carboniferous (Pennsylvanian), 2,300–2,800 feet:
- Dunkard, or Permo-Carboniferous Series, 1,100–1,200 feet
- Monongahela Series, 260–400 feet
- Conemaugh Series, 500–600 feet
- Allegheny Series, 225–350 feet
- Pottsville Series, 250–300 feet

Lower Carboniferous (Mississippian), 450–1,000 feet:
- Mauch Chunk Shales, 40–250 feet
- Greenbrier Limestone, 15–150 feet
- Pocono Sandstones, 400–600 feet

Devonian:
- Catskill Sandstones, 300–500 feet
- Chemung and Hamilton Shales, penetrated in Wheeling deep well nearly 2,000 feet without reaching the Corniferous.

The Dunkard beds are the so-called "Barren Measures," but they carry several seams of coal; these are unimportant commercially but

they are useful for purposes of stratigraphical correlation. The Monongahela series carries the famous "Pittsburgh" coal bed. It is from 8 to 10 feet thick, mined from nearly 60 mines in this one district. The "Sewickley" is generally about 6 feet thick; it is second only to the Pittsburgh coal in economic importance. In the Allegheny series the most valuable coal seams are the Upper and the Lower Kittanning.

The Pocono series consists of alternating beds of sandstones and brown shales. It has been of enormous economic importance, because it contains four oil sands: the Keener, Big Injun, Squaw, and Berea. Oil and gas have been produced from various sands between the Moundsville in middle Conemaugh series down to the Bayard sand near the bottom of the Catskill beds; this is a vertical range of about 1,000 feet.

The report is accompanied by a set of three maps, to make plain the topography, the soils, and the economic geology of the district. The structure contours on the geological map are of especial value and significance in relation to the deposits of coal, gas, and oil.

Cabell, Wayne, and Lincoln counties are in the extreme western part of the state. Their topography is very rough, typically mature. The geological formations dip in a general northerly direction, but they have been somewhat warped by minor folding. The following formations outcrop in these counties, in the order of their appearance, starting from the north: the Dunkard, Monongahela, Conemaugh, Allegheny, and the Kanawha. The removal by erosion of the Monongahela from most of this area, and the irregularity and distribution of the Pittsburgh seam, where the formation still remains, prevent this district from being a great coal-producer. The coal of the Allegheny and Kanawha series is not enough to put these counties very far from the smallest coal-producers in the state. They produce about 100,000 tons per annum as compared with about 4,500,000 tons shipped from the mines of the other three counties reported upon.

The north-central part of these counties is a large oil-producing region; the report suggests that the southern part of the area will be a fruitful field for the oil and gas prospector.

For each of these counties there is a set of three maps which show the agricultural soils, the topography, and the general and economic geology. The maps are made on the scale of one inch to the mile; they have structure contours showing the depth below the surface of the economic deposits and the stratigraphical position of oil and gas accumulations.

T.T.Q.

Water Reptiles of the Past and Present. By S. W. Williston, of the University of Chicago. Chicago: The University of Chicago Press, 1914. Pp. 251, text fig. 131.

It is a deplorable fact, but nevertheless true, that the popular knowledge of present-day reptilian life is very limited. In the opinion of the author of this volume: "In most persons the word reptile incites only feelings of disgust and abhorrence; to many it means a serpent, a cold, gliding, treacherous, and venomous creature shunning sunlight and always ready to poison."

Naturally the public's knowledge of extinct reptiles is much more fragmentary and usually confined to inflated newspaper accounts of "monster" dinosaurs, and a few other forms, possibly. In this volume the author attempts to make the reader familiar with one of the most interesting phases of reptilian life, perhaps, the aquatic forms and the modifications that fitted them for this mode of life. Chaps. i–vi define, in an unpretentious manner, the Reptilia and tell of their occurrence in the rocks and their collection and restoration, their anatomy and classification, their distribution geographically and geologically, and the laws some of them have followed in their adaptation to a life in the water. A classification is given in chap. ii that conforms in the main with the ideas of the more conservative paleontologists. The class is divided into the following orders: the Cotylosauria, Chelonia, Theromorpha, Therapsida, Sauropterygia, Icthyosauria, Squamata, Rhynchocephalia, Parasuchia, Crocodilia, Dinosauria, and Pterosauria. To these the Proganosauria, the Protorosauria, and the Thalattosauria are added and are provisionally given ordinal rank. Chaps. vi–xvi take up in order the Sauropterygia, Anomodontia, Ichthyosauria, Proganosauria, Protorosauria, Squamata, Thalattosauria, Rhynchocephalia, Parasuchia, Crocodilia, and the Chelonia. The various changes these forms have undergone in their adaptation to a life in water, their habits, and, in many cases, interesting bits of history connected with the discovery of the specimens, are told in a fascinating manner.

A few inaccuracies occur in the text and the lettering of some of the figures and some minor additions might be suggested. Among the vertebrate localities given on p. 52 should be included the Pennsylvanian beds from which Case has described vertebrates (*Ann. Carnegie Mus.*, IV [1908]), and on p. 54 might be mentioned the Hallopus beds of Marsh in connection with the Lower Jurassic.

One of the noteworthy features of the book is the large number of excellent illustrations which are, for the most part, the work of the

author. Especially interesting are the life-restorations, some of which are here published for the first time. There is no one better fitted to discuss the subject-matter contained in this book than the author of this volume. In it he gives to the public the benefit of his observations gathered from over forty years of actual experience in the reptilian field. It is rare indeed that a subject is given such an authoritative, scientifically exact treatment combined with a style so thoroughly understandable and interesting to the non-scientific reader. The work is sure to be very popular with the scientist and the general public alike.

M. G. MEHL

The Climatic Factor as Illustrated in Arid America. By ELLSWORTH HUNTINGTON, Assistant Professor of Geography in Yale University, with contributions by CHARLES SCHUCHERT, ANDREW E. DOUGLASS, and CHARLES J. KULLMER. Carnegie Institution of Washington, Publication No. 192, 1914. Pp. vi +341, plates 12, maps 2, text figs. 90.

This volume has bearings which make it important to the geographer, historian, archeologist, meteorologist, and geologist, occupying a field where all these sciences meet, but in this review the volume will be discussed from the geological point of view only.

The purpose of the work is to determine the degree to which climatic changes have taken place in southwestern America during the past 2,000 to 3,000 years. In arid and semi-arid regions the amount of rainfall, as affected by pulsatory changes of climate, becomes most variable and critical.

In addition to the study of the climatic changes shown by the expansion and restriction of ancient peoples in America, as controlled by changes in water-supply or vegetation, the present volume contains two novel lines of attack. The first of these is the use of river terraces as evidences of minor climatic changes occurring within the past few centuries as well as in the more distant past. The second is the measurement of the growth rings of trees. Professor A. E. Douglass gives an introductory chapter on a method of estimating rainfall by the growth of trees. He shows that the rings vary in thickness and correlates the rate of growth with the records of rainfall. Following this, Huntington enters upon a most interesting discussion of the curve of growth of the giant redwoods of California. The data were obtained by careful measurements from stumps and extend back with a large number of trees as much as 2,000 years, with a few trees to 3,000 years. The geological importance of this work is readily seen. As Lyell showed that the present is the key to the past in the crustal history of the earth, similarly the key to the climatic history is to be found in the study of the present climates and

their fluctuations. It is found that the changes, though moderate in amount, are more or less sudden and pulsatory. In the search for causes Huntington consequently assigns crustal movements and changes in atmospheric composition as the broad factors of ultimate control, but the pulsations which appear within the historic record and which extend beyond in the record of moraines left by the oscillations of glaciers, the strands of salt lakes, and terraces made by river action cannot, he argues, be due to these causes. Variations in solar radiation are assigned as the most probable cause. Rhythms measured by tens of miles marked the retreat of the Pleistocene ice sheets; rhythms measured by inches, by feet, and by tens of feet are found in the sediments of many geological formations. In many cases they can hardly be ascribed to crustal causes; they are too many and both too long and too short to fit the precession cycle. Thus the geological record is suggestive that our sun through all of terrestrial history has been a variable star.

The second part of the volume consists of a chapter entitled "The Climates of Geologic Time, and is by Professor Schuchert. There is assembled in thirty pages an account of the various lines of evidence which indicate geologic changes in climate. These are finally correlated in a single chart. The curves of coal-making, limestone-making, aridity, and temperature are given, together with curves showing the movements of the strand line and epochs of diastrophism. While the curves are of course only of qualitative value, they serve to show the variability and the cyclic nature of all these factors through geologic time. This chapter thus gives on a large scale and in distant perspective what the first part of the book gives in minute scale and for the human present.

It would appear that the work of Douglass and Huntington on tree growth opens up a field which deserves further study; a study which should be prosecuted within a few years. In this reconnaissance Huntington has averaged together the measurements of many individual trees. But these have grown under unlike conditions of altitude, slope exposure, and ground water. The averaging of these unlike conditions has tended toward obscuring the amount of short climatic oscillations and the trend of longer changes. An intensive study of stumps selected with respect to these variable controls and an exact dating of special sequences of rings by comparison of stumps would seem to be the next step. But in the meantime decay is blurring year by year this most valuable record.

There are of course in this volume degrees of emphasis and points of view which could be questioned, but within the limits of a short review it would confer a wrong emphasis to single out any point for critical comment, when the book as a whole is a contribution of the first order, in facts, in ideas, and in completeness of presentation. It adds fundamentally to the science which in the future will be named if not now—paleoclimatology.

J. B.

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THE

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FEBRUARY-MARCH 1915

POST-CRETACEOUS HISTORY OF THE MOUNTAINS OF CENTRAL WESTERN WYOMING[1]

ELIOT BLACKWELDER
University of Wisconsin

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PART I

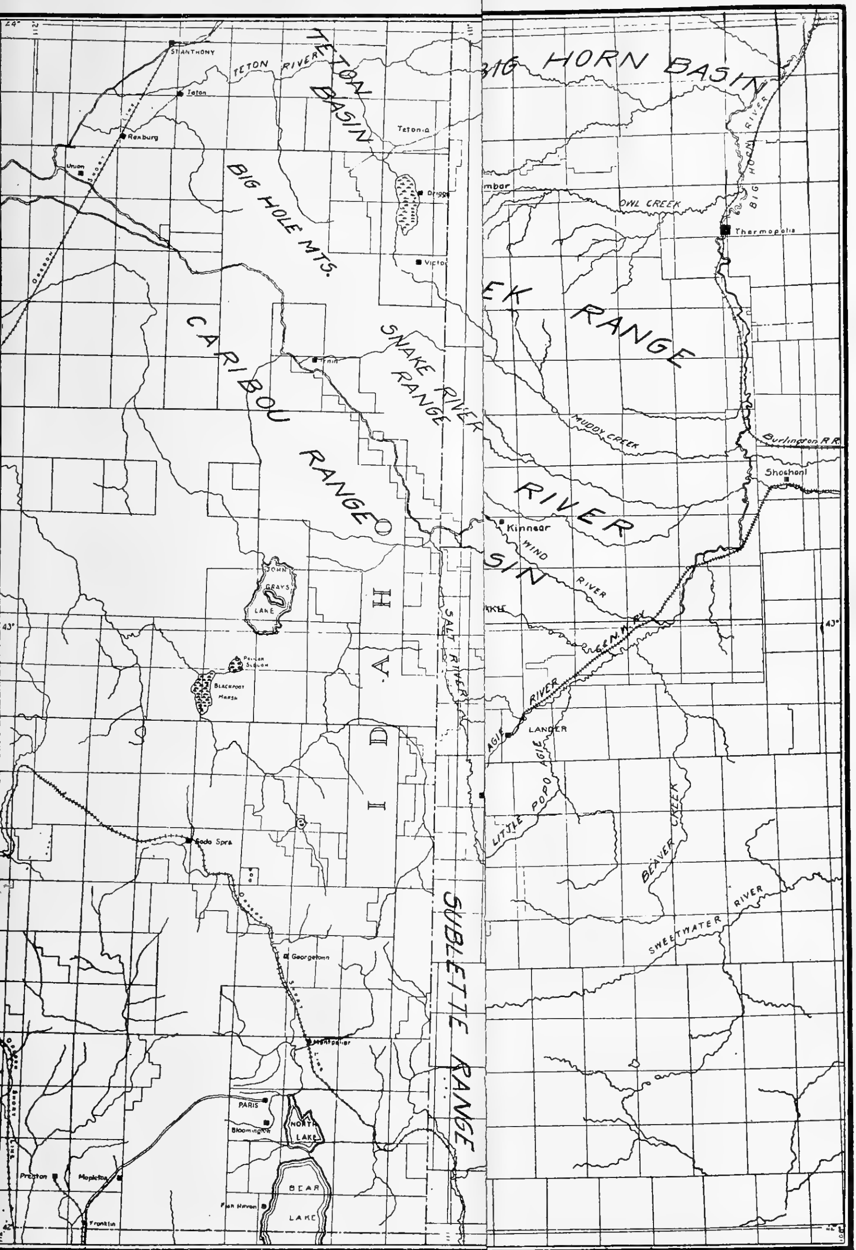
INTRODUCTION

Subject and scope.—In this essay it is my purpose to discuss the geologic history of a part of western Wyoming, from the close of the Cretaceous period to the present. The field under consideration is limited naturally by the district which it has been my privilege to explore. For the interpretation of Cenozoic history the district is especially favorable because that era is represented by deposits of several ages, and the physiographic conditions are so varied that they afford the means of interpreting many of the later events. It is scarcely necessary to state that many of the problems which arose in the course of the study remain unsolved or but partly solved, for such is the nature of most complex questions.

Data and acknowledgments.—The data for this work have been obtained largely from my own field examination of the district for the United States Geological Survey during the summers of 1910, 1911, 1912, and 1913. In the summer of 1910 a somewhat rapid reconnaissance trip was made from Montpelier, Idaho, through Jackson Hole and the Wind River basin to Thermopolis, Wyoming. In 1911 we ascended Green River from the Union Pacific Railroad, made a brief examination of the southern part of the Wyoming Range, carefully studied the Fall River basin and the headwaters of Green River, and devoted the remainder of the season to an examination of the Gros Ventre Range and the highlands north of it. In 1912, a detailed study was made of the west slope of the Teton Range near the Idaho line. The summer of 1913 was devoted to the Owl Creek Range and the Wind River basin and Range. In this field work I have had the assistance of Messrs. J. M. Jessup, C. W. Tomlinson, Hyrum Schneider, and D. Dale Condit.

Earlier seasons in surrounding districts afforded me an opportunity to gather information which has value by way of comparison with western Wyoming. Thus the Bighorn Mountains in north-central Wyoming were studied in 1902, the Laramie Range in the southeastern part of the state in 1907–8, and the Wasatch Range of northern Utah in 1909.

In addition, some facts and many interpretative suggestions have been obtained from the writings of other students of the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SNAKE RIVER RANGE
BIG HORN BASIN
OWL CREEK
BIG HORN RIVER
Thermopolis
EK RANGE
MUDDY CREEK
Burlington R.R.
Shoshoni
RIVER
SIN
Kinnear
WIND RIVER
JOHN GRAYS LAKE
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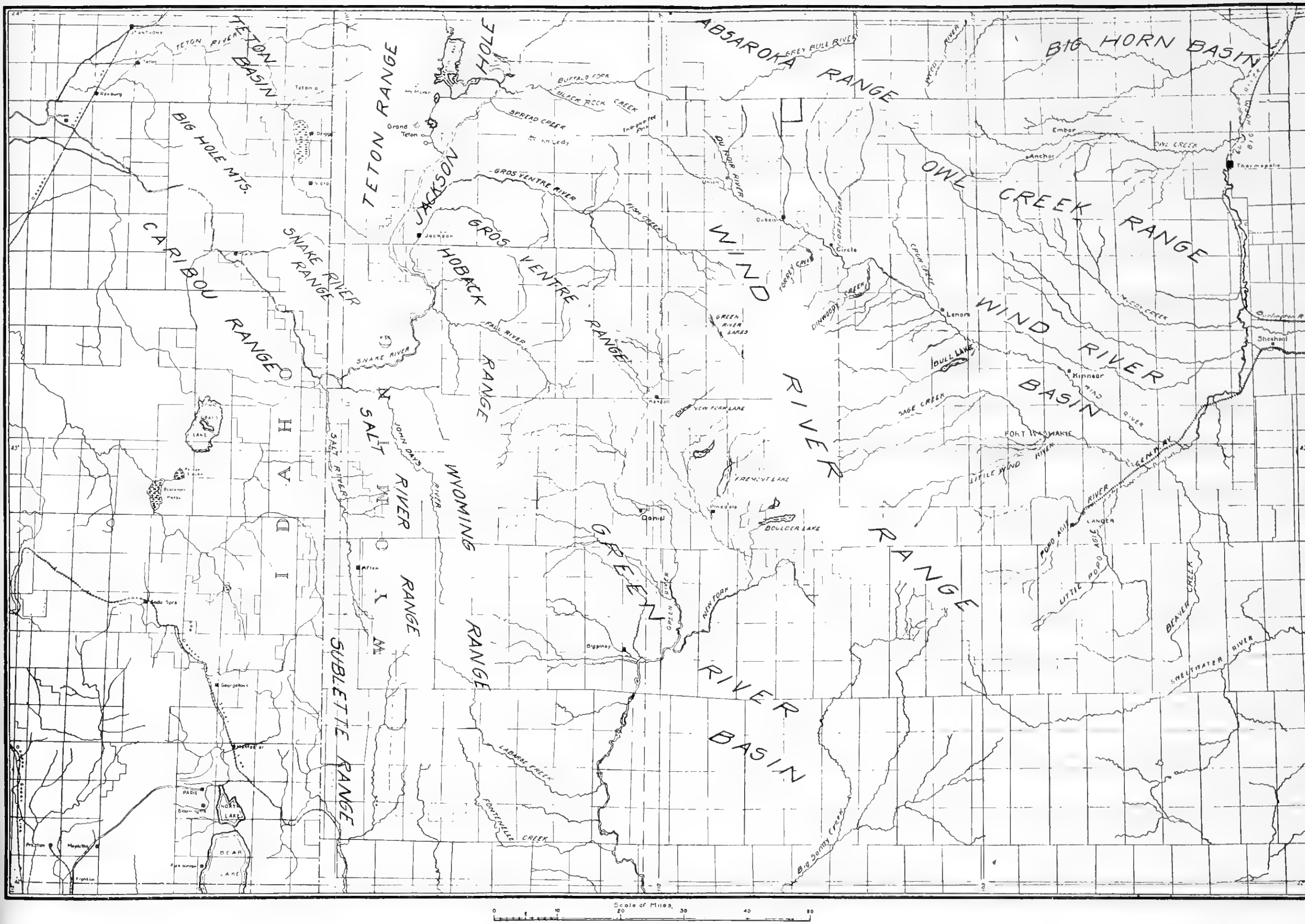
I D A H
SALT RIVER
43°
C.&N.W.RY.
RIVER
AGIE
LANDER
LITTLE POPO AGIE
BEAVER CREEK
SWEETWATER RIVER
Soda Sprs
Georgetown
Montpelier
SUBLETTE RANGE
PARIS
Bloomington
NORTH LAKE
BEAR LAKE
Fish Haven
Preston
Mapleton
Franklin



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TETON RANGE
JACKSON HOLE
ABSAROKA RANGE
GREY BULL RIVER
BIG HORN BASIN
BIG HOLE MTS.
BUFFALO FORK
BLACK ROCK CREEK
SPREAD CREEK
GROS VENTRE RIVER
JACKSON
GROS VENTRE RANGE
HOBACK RANGE
CARIBOU RANGE
SNAKE RIVER RANGE
SNAKE RIVER
OWL CREEK RANGE
WIND RIVER RANGE
WIND RIVER BASIN
GREEN RIVER LAKES
BULL LAKE
SAGE CREEK
FORT WASHAKIE
LITTLE WIND RIVER
NEW FORK LAKE
FREMONT LAKE
BOULDER LAKE
SALT RIVER
SALT RIVER RANGE
JOHN DAYS RIVER
WYOMING RANGE
IDAHO
WYOMING
GREEN RIVER BASIN
GREEN RIVER
NEW FORK
POPO AGIE RIVER
LITTLE POPO AGIE
BEAVER CREEK
SWEETWATER RIVER
SUBLETTE RANGE
LABARGE CREEK
FONTENELLE CREEK
BIG SANDY CREEK
BEAR LAKE
NORTH LAKE
Jackson
Grand Teton
Circle
Lenore
Kinnear
Shoshoni
Thermopolis
Embar
Anchor
Burlington R.R.
Pinedale
Daniel
Big Piney
Afton
Soda Spgs
Georgetown
Paris
Preston
Lander
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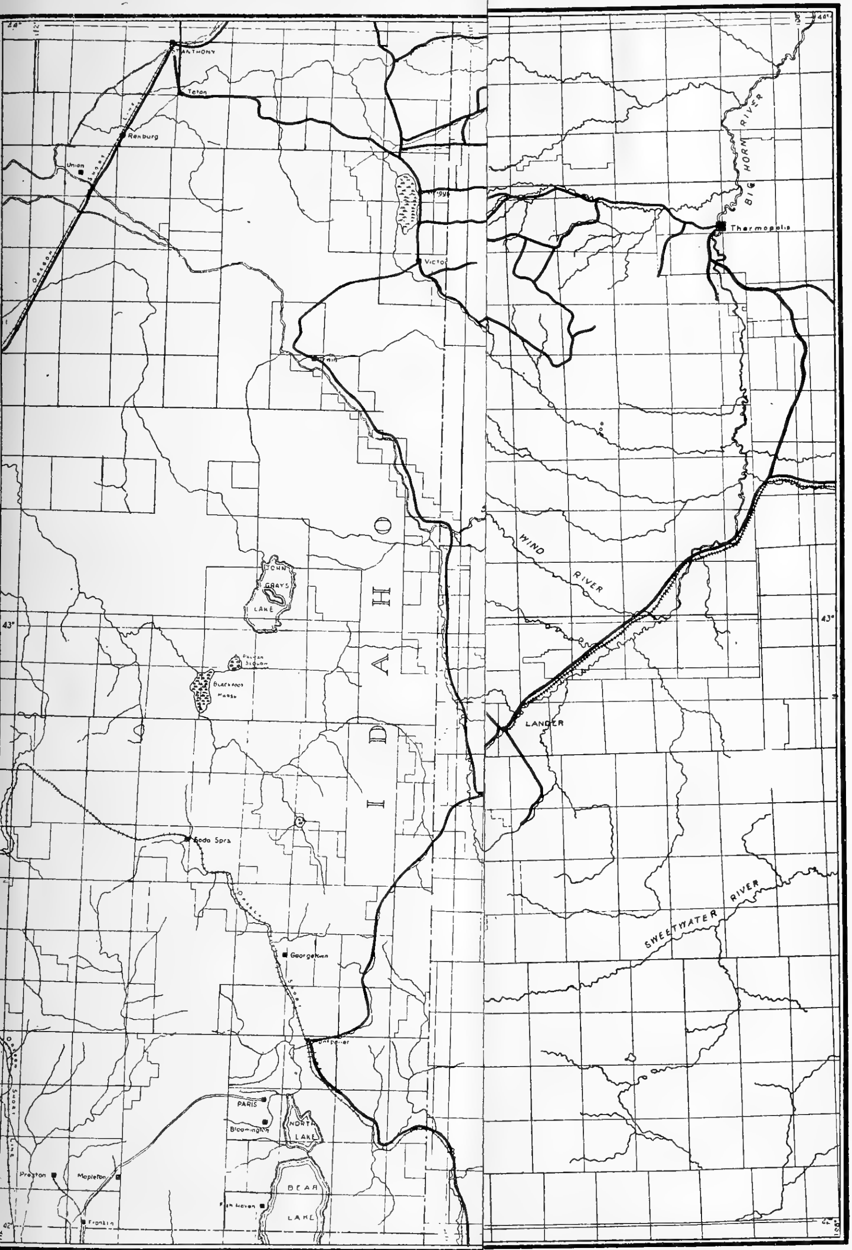
0
10
20
30
40
50

FIG. 1



ANTHONY
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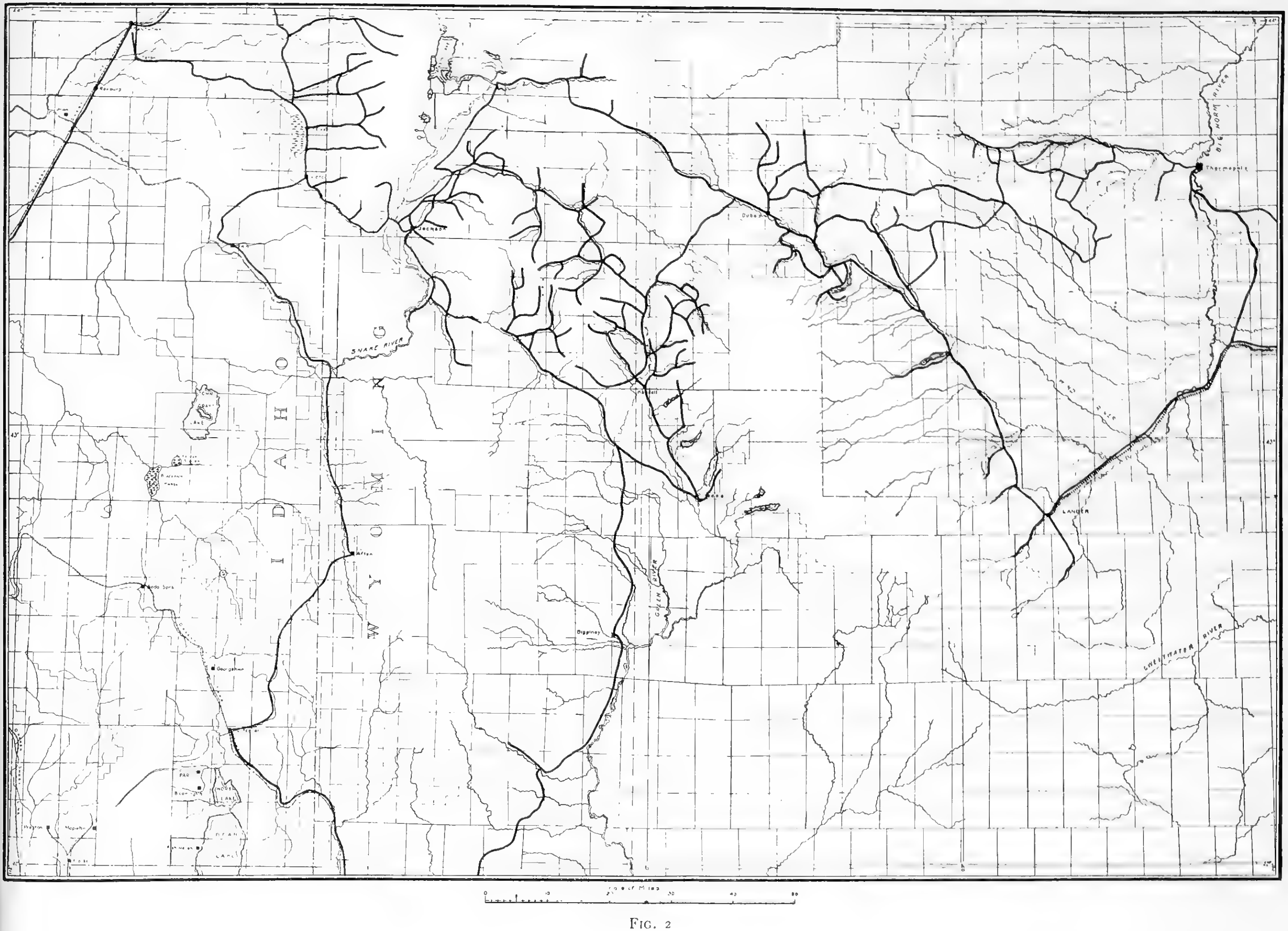
Rexburg
Union
Victor
Irwin
JOHN GRAYS LAKE
Blackfoot Marsh
Soda Sprs
Georgetown
Montpelier
PARIS
Bloomington
NORTH LAKE
BEAR LAKE
Fish Haven
Preston
Mapleton
Franklin
I D A H O
BIG HORN RIVER
Thermopolis
WIND RIVER
LANDER
SWEETWATER RIVER
43°



SNAKE RIVER
IDAHO
WYOMING
Jackson
Dubois
Thermopolis
BIG HORN RIVER
Lander
Afton
GREEN RIVER
SWEETWATER RIVER
Big Piney
Georgetown
Soda Spr.
Preston
BEAR LAKE
Rexburg

FIG. 2

Rocky Mountains, and especially from papers on districts outside the one immediately considered.

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1. Fremont, Lieutenant John C. On the expedition to the Rocky Mountains in 1841. A general account of the Wind River Range and the adjacent Wyoming plateaus, but there is very little strictly geologic information.

B. TERRITORIAL SURVEYS

1. U.S. Geological and Geographical Survey of the Territories ("Hayden Survey"), 1872–79. The geological exploring parties in immediate charge of Bradley, Comstock, St. John, and Endlich, and under the general supervision of F. V. Hayden, made the first thoroughgoing reconnaissance of western Wyoming and parts of adjacent states. Considering the dangerous and unsettled character of the country, the lack of previous knowledge, and the short time that could be devoted to each district, the results obtained by these parties are remarkable and deserve the highest praise. The work of Orestes St. John I have found especially trustworthy. The members of these parties made a fairly good topographic map of the district and reinforced it with perspective sketches which not only take high rank for their artistic merit, but are geologically accurate. They correctly ascertained the long sequence of rock formations and interpreted the structure of the ranges with but few mistakes. In their voluminous reports one may find many suggestions about the geologic history of the district here considered, and although some of these suggestions have not been sustained by later studies, many of them will probably stand indefinitely. Altogether, the Hayden reports contain the only comprehensive and detailed account of the geology of western Wyoming that has been written.

C. SHORTER PAPERS ON SPECIAL QUESTIONS OR ON SMALL PARTS OF THE DISTRICT

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2. Weeks, F. B. "A Reconnaissance in Jackson Basin," abstract in *Science*, N.S., IX (1899), 454.
3. Woodruff, E. G. "The Lander Coal Field," *U.S. Geol. Survey, Bull. 316*, 1907, pp. 242–43. A mere note.

[1] Only such papers are here listed as have a direct bearing either upon the geology of the district in question, or upon the problems of the Cenozoic era in adjacent parts of the Rocky Mountain region.

4. Baker, C. L. A report, thus far unpublished, on the west end of the Wind River Range. In this paper there is described for the first time the Wind River peneplain. It doubtless contains many other data relating to the subject of this thesis (work done in 1908–9).
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2. Hague, Iddings, Weed, and others. "Geologic History of the Yellowstone National Park," *Smiths. Inst., Ann. Rept.*, 1891–92, pp. 133–51.
3. ———. "The Age of the Igneous Rocks of the Yellowstone National Park," *Am. Jour. Sci.*, 4th Ser., II (1896), 445–57.
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12. Sinclair, W. J., and Granger, W. "Eocene and Oligocene of the Wind River and Bighorn Basins," *Bull. Am. Mus. Nat. Hist.*, XXX (1911), 83–117.
13. ———. "Notes on the Tertiary Deposits of the Bighorn Basin," *ibid.*, XXXI (1912), 57–67.
14. Sinclair, W. J. "Some Glacial Deposits East of Cody, Wyoming, and Their Relations to the Pleistocene Erosional History of the Rocky Mountain Region," *Bull. Geol. Soc. Am.*, XXIII (1912), 731 ff.
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POST-CRETACEOUS GEOLOGIC HISTORY

Antecedent conditions.—In order to understand the effects produced by the various events and changes after the Cretaceous period, the reader should recall the general character of events before that time and particularly the conditions which had been brought about at the close of the Mesozoic era. These will be outlined without evidence or argument.

The oldest rocks of the district, generally referred to the Archean system, were folded and metamorphosed, and subsequently worn down to a peneplain before the middle of the Cambrian period. Beginning with the Cambrian and extending on through all of both Paleozoic and Mesozoic eras, sediments were spread over this peneplained surface until it had been buried to a depth of many thousand feet. Most of these sediments were deposited in the sea, but noteworthy portions were laid down upon land. Occasionally there were short episodes in which the region was subject to erosion, but these periods were not accompanied by orogenic disturbances and served to reduce the strata but little. The variety of sedimentary rocks thus produced is shown in the accompanying columnar section, and the importance of the varying erodability of the different beds will be better appreciated when the discussion of the modern physiographic forms is undertaken.

Since the latest and thickest of these deposits are partly marine and extend more or less uniformly over the entire district—except

where subsequently eroded from the crests of the anticlines—it is tolerably safe to assume that western Wyoming, near the close of the Cretaceous period, was a nearly level plain of aggradation very near sea-level, and wholly devoid of outcrops of the Paleozoic and older rocks (Fig. 3). It is deemed unnecessary to apologize here for disregarding the ancient view that each area of Archean outcrops has been an island ever since the Cambrian.

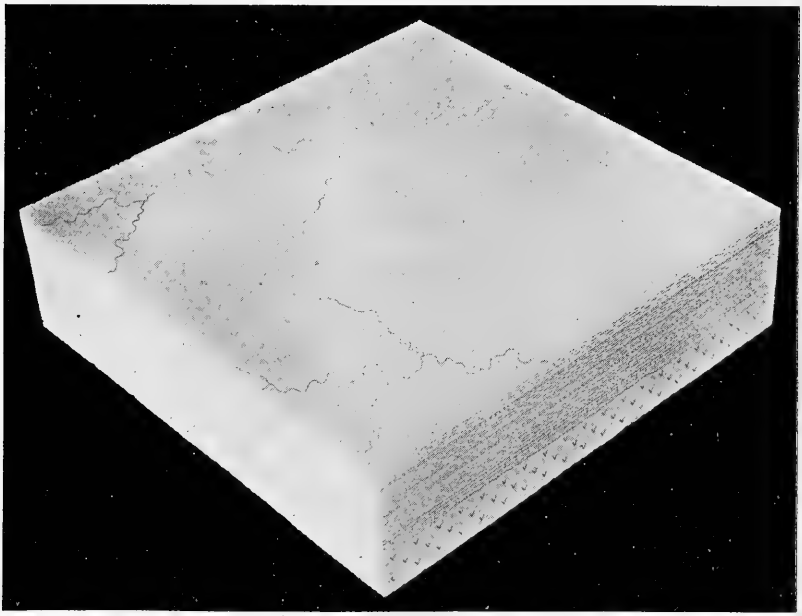

Fig. 3

Post-Cretaceous orogenic disturbance.—The Cretaceous and underlying strata have been compressed into a series of folds which generally trend northwestward through central Wyoming, and bend around to the south in the southwest part of the state. The prevailing structures in the district are open anticlines and synclines oversteepened on the southwest sides. Some of the largest folds, such as the Wind River and Gros Ventre anticlines, broke along their southwest flanks and formed overthrusts of large displacement. The most intense folding in this district generally is to be found on

the lower sides of these overthrusts. South and southwest of Jackson Hole most of the folds are much more closely appressed, and the strata generally stand upright. It is a fact worth noting that this deformation was apparently not accompanied by volcanic activity within the region under consideration. I have attempted to

TABLE I

CONDENSED TABLE OF FORMATIONS

Age	Character	Thickness in Feet	Description
Quaternary	Superficial deposits	1–500	Glacial drift, talus, landslides and earth-flows, river gravel and silt, loess, and residual soils
Tertiary (slightly disturbed)	Volcanics	Over 4,000	Stratified agglomerate and tuff with a few lava flows
	Wind River formation and associated strata	1,500 to 6,000	Largely clay and soft sandstone with local conglomerates and lava beds
Mesozoic (folded)	Shale and sandstone	6,000 to 10,000	Largely shale and clay with many beds of sandstone of which a few are thick and massive. Occasional thin layers of limestone, coal, and chert
Paleozoic (folded)	Limestone, etc.	2,000 to 3,500	Largely limestone or dolomite, some beds very massive and thick. Interbedded with shale, thin limestone, sandstone, and chert
Archean (?)	Metamorphic and igneous	Indefinite	Massive gneisses and some schists, with granitoid intrusions

show by means of a stereogram (Fig. 4) the general character of the folds and faults produced at this time as they might have appeared if they had been completed without being eroded. The effects of subsequent warping, faulting, and erosion have been eliminated. It is probable that there are more folds in the district than are shown on this model, for in other parts of the Rocky Mountains sharp folds in the Cretaceous rocks are occasionally revealed in canyons which locally cut through the superficial Tertiary cover.

The geologic age of the folding may be roughly ascertained without difficulty, but to determine it with a high degree of accuracy

and to measure its duration are not feasible on the basis of the facts obtained in this district alone. The youngest beds involved in the folding are of the Montana group, the age of which is well established by marine fossils as middle Upper Cretaceous. Still younger strata, which now stand in a nearly vertical attitude in the canyon of Buffalo Fork, have yielded plant remains, identified by Dr. F. H. Knowlton as belonging to the Fort Union flora. Elsewhere in Wyoming it is reported[1] that the Fort Union and other formations which are correlated with it have been folded to the same degree as the underlying marine Cretaceous strata and are overlain in angular unconformity by the Wasatch Eocene and equivalent

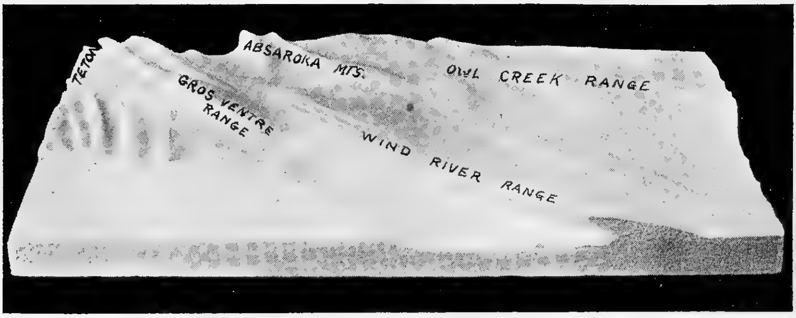

FIG. 4

formations. However, in some of these districts the Cretaceous rocks are decidedly more folded than the Fort Union beds. This has been most clearly worked out in southwestern Wyoming by Veatch and Schultz.[2] The Fort Union formation and contemporaneous deposits elsewhere are generally referred to the "basal Eocene" or Paleocene,[3] although both stratigraphically and pale-

[1] Max W. Ball and E. Stebinger, *U.S. Geol. Survey, Bull. 381*, 1910, p. 194; D. E. Winchester, *U.S. Geol. Survey, Bull. 471*, 1912, p. 47.

[2] A. C. Veatch and A. R. Schultz, "Geography and Geology of a Portion of Southwestern Wyoming," *U.S. Geol. Survey, Prof. Paper 56*, 1907.

[3] Although the term "Paleocene" has not yet been sanctioned by the U.S. Geological Survey and is not in use by all American geologists, it is so much needed in this instance in order to discriminate between "Lower Eocene" and other early Tertiary rocks still older than "Lower Eocene" that it will be freely used in this paper. The term "Eocene" will refer in these pages to those strata now commonly included in "Lower," "Middle," and "Upper Eocene," or, in terms of Rocky Mountain stratigraphy, the sequence from Wasatch to Uinta formations.

ontologically they are rather closely related to the underlying Cretaceous. Following current usage in referring these formations to the base of the Tertiary system, it appears that the principal deformation occurred after the laying-down of the Paleocene series, although in some parts of the Rocky Mountains an important orogenic movement had preceded the Paleocene.

The oldest strata which were not affected by this particular episode of deformation within the limits of the district belong to the early Tertiary series, represented east and southeast of Jackson Hole by the Pinyon conglomerate, in Green River valley by the Green River formation and associated strata, along Wind River by the Wind River formation, and in the Bighorn basin by the so-called "Wasatch" beds. Mammalian and plant fossils found at various points in these districts are said to prove that the early Tertiary strata are Eocene and that the sequence generally, although not everywhere, begins with the Lower Eocene or Wasatch stage. This is in harmony with conditions generally throughout the Rocky Mountains.

The old familiar fact is thus redetermined—that the folding took place between the close of the Cretaceous and the deposition of the Lower Eocene. Assuming that the strata in the canyon of Buffalo Fork, mentioned above, have been correctly referred to the Fort Union formation, it is suggested as probable that the most vigorous folding took place between the Paleocene and Eocene epochs.

Early Eocene erosion.—So nice is the adjustment of stream activities that no sooner did the post-Cretaceous folds commence to be bulged above the grade-level of their time than they were subject to erosion. The fact that they were thus eroded is amply attested by the unconformity at the base of the Eocene (Wasatch, etc.) strata. Since the latter now rest upon the trunkated edges of any and all of the Paleozoic and Mesozoic formations, and in some places even upon the Archean, it is evident that the post-Cretaceous folds were completely trunkated during the early part of the Eocene epoch.

Whether this resulted at any time in bringing the entire region to the condition of a peneplain is still an open question. The

surface which was finally produced just before the deposition of the Eocene strata is, however, preserved in the sub-Eocene unconformity. Profiles of its forms can be observed along the canyon walls in several of the ranges. At some points near the west end of the Owl Creek Mountains the old land surface has recently been exhumed by the denudation of the soft Eocene clays and yet not seriously disfigured during the process (Fig. 5). This surface was post-maturely hilly upon the harder rocks, and locally retained a relief of over 1,000 feet; but, as would be expected from their unresisting character, the softer Cretaceous and Jurassic beds were generally worn down to plains. In many places the

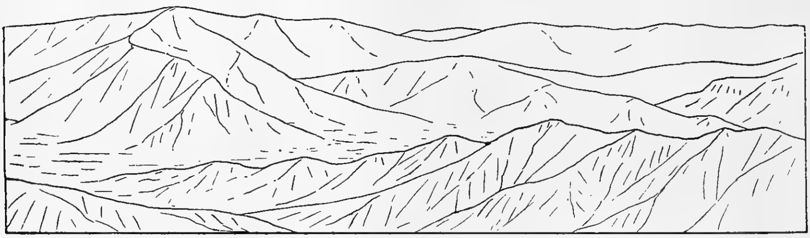

FIG. 5

Eocene now crosses the axes of the anticlines and thus shows that the ranges had been largely worn low.

Deposition of the Eocene and Oligocene formations.—Before the hilly and even mountainous parts of the early Eocene surface could be reduced to base-level, there set in more or less abruptly a change which caused the erosive processes to be supplanted in the depressions by deposition. The nature of the change is not obvious, but it may have been either a warping, which unbalanced the river systems, or a desiccation, which substituted torrential or seasonal erosion and alluviation for the slow continuous degradation of a moister epoch. Since the effects seem to have been nearly alike from Montana to Mexico, a widespread and rather uniform change is suggested. The climatic change answers this requirement, but there is uncertainty as to whether so great a thickness of sediment can be accounted for in this way. There is no question as to the competency of warping to bring about the erosion of some

areas and coincident filling of others. It appears, therefore, to be the more promising factor in this case. That the new state of things continued almost, or quite, without interruption for a long lapse of time is shown by the fact that the eroded surface is now buried in some places by a thickness of apparently conformable sediments which cannot well be less than 4,000 feet thick and may easily be much more. Remnants of these beds are now to be found in all the principal lowlands of the district, such as Green River basin, Jackson Hole, and Wind River valley, and also in the Mount Leidy highlands at the northwest end of the Wind River uplift. The strata vary considerably from place to place, as they might be expected to in view of their terrestrial origin and the irregular surface upon which they were deposited.

The origin of these Tertiary deposits is not wholly obvious, and their designation as "lake-beds" by the Hayden Survey is open to grave doubt. It is safe to say that they are not marine, for instead of marine fossils they contain the remains of land plants and land mammals, as well as of freshwater fishes and mollusks. As working hypotheses we may suppose that they could have been deposited (*a*) by lakes and marshes, (*b*) by graded streams upon their floodplains, (*c*) by wet-weather streams making alluvial fans, or (*d*) by the wind upon dry land surfaces. It is not inherently improbable that each of these agencies may have played an important part. The problem compels more analysis.

The Pinyon conglomerate (Fig. 6) of the Mount Leidy highlands has already been mentioned as highly suggestive of the work of powerful aggrading streams with constantly shifting channels. It appears to be identical with deposits being made today by those means, and no other agency seems competent to produce it. The pebbles are nearly all quartzose in composition and are in large part foreign to the district. This suggests that they have been transported far, and that they are the residue of long-continued attrition. If this be granted, it may be taken to indicate that the Pinyon conglomerate was deposited largely by aggrading rivers of considerable length and power, rather than chiefly by local creeks from the surrounding mountains.

Along the flanks of the Owl Creek and Wind River mountain ranges the basal part of the Tertiary strata contains wedge-shaped beds of coarse conglomerate interleaved with the finer sediments and pinching out away from the mountains. The pebbles are ill rounded and consist wholly of rocks which outcrop near by. The lithologic varieties are many and not well assorted. These facts suggest the local deposition of alluvial fans by mountain creeks along the border of a lowland. Similar conditions exist today in the California valley.

FIG. 6

Again, some small parts of the sequence have evidently been deposited in lakes, since from the valley of the north fork of the Gros Ventre River, Mr. Perry[1] of the Hayden Survey reports five successive beds of freshwater limestone containing recognizable lacustrine shells. There is equally good evidence that some of the material was laid down in marshes, for thin lignitic seams have been found in many localities, and especially in the section last mentioned. There Perry found in a single exposure no less than forty-seven beds of lignite, most of them very thin. Similar beds occur in the Bighorn and Wind River basins.

[1] Hayden Survey report for 1878, Part 2, pp. 223–24.

The banded clays, usually of pink and gray colors, which constitute the bulk of the Tertiary beds along the head of the Gros Ventre River, throughout the Wind River basin, and in the Bighorn basin, suggest a different mode of origin. Sinclair,[1] from a careful study of these clays and their associated sandy beds in the Bighorn basin, reaches the conclusion that they are river floodplain deposits laid down under a moderately dry climate. The color banding he ascribes to alternating cycles of moist and dry climate (pp. 116–17). So regular is the repetition of their red and gray layers that an oscillating control is strongly indicated. Loomis[2] also reached the conclusion that the clays and silts of the Wasatch formation in the Bighorn basin were floodplain deposits. In this connection he called attention to the fact that, of the very large number of fossils which had been found in the beds, only 10 per cent are aquatic, whereas the most abundant forms are horses and other animals which frequent the relatively dry plains. Even the aquatic animals were turtles, crocodiles, and fishes, all of which inhabit rivers.

Such observations as I have made tend to confirm this hypothesis of floodplain deposition, under subarid or steppe conditions. The more sandy layers consist of small angular fragments of quartz, prisms and slivers of hornblende, hexagonal plates of mica, and sharp bits of feldspar. No volcanic ash was observed in the typical striped Wind River beds. The undecayed condition of the grains precludes a warm moist climate, and the prevalent reddish or pinkish color is evidence of thorough oxidation, which would have been inhibited in a cold moist climate. The general lack of abrasion suggests that the particles have not been transported very far, or at least were not worked over continuously for a long space of time. To Loomis and Sinclair's hypothesis I may add the suggestion that the Eocene sediments probably originated as regolith due to rock disintegration and eolian abrasion in a dry climate. Much of the angular quartz and feldspar dust may well be the product of sandblast action upon rock outcrops. The regolith, once formed, was

[1] W. J. Sinclair and W. Granger, "Eocene and Oligocene of the Wind River and Bighorn Basins," *Am. Mus. Nat. Hist. Bull.*, XXX (1911), 83–117.

[2] F. B. Loomis, "Origin of the Wasatch Deposits," *Am. Jour. Sci.*, XXIII (1907), 362.

probably washed down from its points of origin by temporary hill streams, sifted from the coarser fragments, imperfectly sorted mineralogically, and soon deposited on the flat bottoms of broad, intermontane basins. It is a suggestive fact that the sandy silt taken from a cut bank in the broad floodplain of Owl Creek has the same structural, textural, and mineral characteristics as some of the Wind River strata. This silt is a modern floodplain deposit in a drainage basin which does not contain outcrops of the typical Wind River strata. A mere reworking of the Eocene material in this instance is scarcely possible.

Some parts of the Tertiary sequence are obviously volcanic in origin. Thus, along the east side of Jackson Hole, there are considerable flows of andesite and rhyolite in the lower part of the formation. If we may rely upon lithologic similarity and field relations, we may correlate with these the flows which cloak the west slope of the Teton Range, although the relationship is not yet firmly established. At the northwest end of the Wind River Range, where it articulates with the mountains of Yellowstone Park, thick beds of volcanic ash and agglomerate with interbedded glassy lava flows rest upon the pre-Tertiary folded rocks, but are themselves younger than the Wind River Eocene. Traced eastward to Horse Creek, the Washakee Needles, and the valley of Owl Creek, this thick volcanic series is found to rest conformably upon the striped clays of the Wind River formation, with which they intergrade through gray, plant-bearing shales and greenish volcanic sandstones containing petrified logs. A closer examination of the volcanic beds shows that some of them are massive agglomerates, devoid of stratification, whereas other beds are distinctly stratified, cross-bedded, and occasionally interrupted by lenticular sheets of coarse gravel, suggestive of stream channels. The conditions indicated are those which would be found upon low gradient river plains adjacent to active volcanoes.

In order to place these conditions as closely as possible in chronology, it would be necessary to know the ages of all parts of this Tertiary sequence. Unfortunately, this is not fully understood. On the basis of freshwater shells, and land plant leaves which were found in the valley of the Gros Ventre River by the

members of the Hayden Survey, the strata were tentatively correlated with the Wasatch of Utah. Fossil mammals from the Wind River basin[1] have recently been correlated with those of the Wind River and Uinta formations of the Eocene and with those of Lower Oligocene. In the southwest part of the Bighorn basin abundant mammals belonging to the typical Wasatch have been found by Wortmann, Loomis, Granger, and others. The beds in which they occur are so closely similar to those in the Upper Wind River valley and near the head of the Gros Ventre River that the intercorrelation of all three is strongly suggested. Since the volcanic Tertiary beds lie almost entirely at higher elevations, and, where both formations are present, at stratigraphically higher horizons than the Wind River formation, they are probably not older than Upper Eocene. Sinclair and Granger have found Lower Oligocene fossils in beds of tuff and agglomerate east of Lander. Older Tertiary beds in that vicinity do not contain volcanic materials, except for two or three beds of ash in the Middle or Upper Eocene.

In review, the evidence relating to the Tertiary sediments seems to indicate that, from the Wasatch or Lower Eocene epoch until at least the early Oligocene, west-central Wyoming was subject to a semi-arid climate in which there were, however, considerable fluctuations; that streams made widespread deposits in the lowlands; that these were largely of clay, silt, and sand but locally of gravel; and that the wind was a modifying rather than a controlling agency in the process; that lakes and marshes existed locally from time to time; and that volcanic eruptions in the western part of the district, and late in the sedimentary interval, scattered ash and cinders over much of the district. No matter what hypothesis of the origin of the Tertiary beds is adopted, it leads to the conclusion that by the Oligocene epoch the processes of sedimentation had produced one or more broad flat or gently graded plains beneath which were buried deeply the old Eocene hills and many of the lower mountains. There seems to be no means of knowing whether all the ranges were buried by the Tertiary deposits. They now cover

[1] W. J. Sinclair and W. Granger, "Eocene and Oligocene of the Wind River and Bighorn Basins," *Bull. Am. Mus. Nat. Hist.*, XXX (1911), 83–117.

both ends of the Wind River and Owl Creek anticlines and have evidently been eroded from large areas which they once covered. On the other hand, there seems to be no good reason to suppose that some of the principal mountain ranges did not rise above the aggraded plains somewhat as they now do in the Bonneville Lake basin of Utah (Fig. 7).

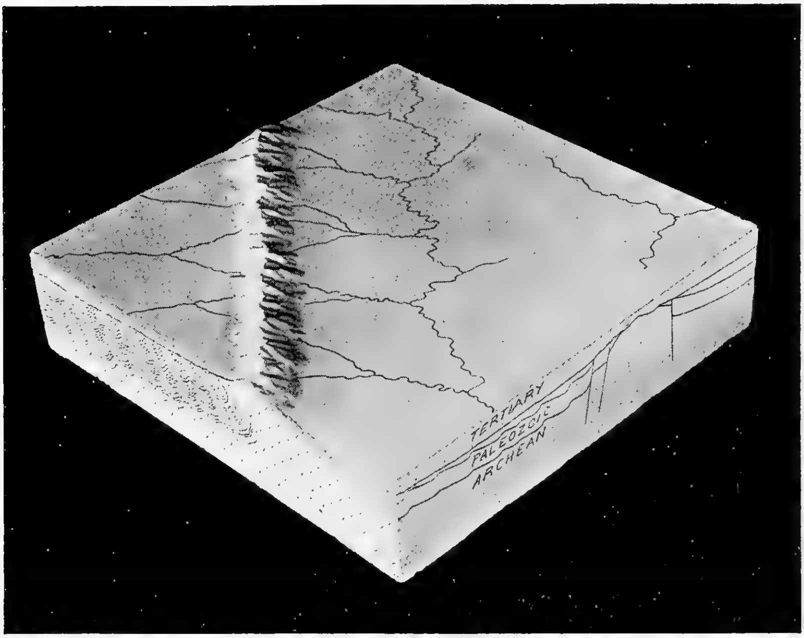

Fig. 7

Mid-Tertiary deformation.—The early Tertiary sediments no longer lie in their original almost horizontal attitude. Nearly everywhere they have been tilted gently, and in a few places bent into folds and broken by faults of large displacement. In Jackson Hole the beds of travertine or freshwater limestone with lava flows dip 10–15° toward the west. Near the forks of the Gros Ventre River the alternating sandstones and clays dip 12–14° northeastward. Near Dubois in the Wind River basin, the finely laminated Eocene clays dip 5–8° away from the Wind River Range. In the southwest part of the Bighorn basin, Sinclair and Granger observed

an anticline in the Wasatch deposits, the dips being 5–7° on either limb. In the slope just east of the E. A. ranch, north of Dubois, there is a fold the southern limb of which dips at an angle of about 35°. The strongest folding I have yet seen in the Wind River beds was found on the north fork of Wind River in the southwest corner of the Kirwin quadrangle (see Fig. 8). There a relatively sharp anticline between two broad flattish synclines exhibits dips of from 5° to over 50°. Although in a few cases the dips may be initial, it is improbable that the fine clays were deposited at angles of even 5–10°.

Faulting is more common than folding in the Eocene strata. Near the mouth of Dinwoody Creek on the north side of the upper

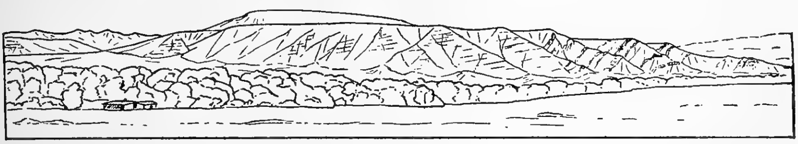

Fig. 8

course of Wind River, a fault of several hundred feet displacement interrupts the nearly horizontal Wind River beds (Fig. 9). On the northwest side of the Teton Range a normal fault of at least 1,500 feet displacement drops the Tertiary volcanic breccias and rhyolitic flows down against Paleozoic strata. At Granite Falls, and again along Jack Creek, both on the south side of the Gros Ventre Range, a coarse conglomerate, apparently the same as the Pinyon formation farther north, has been faulted down against the Paleozoic rocks (Fig. 10). In the southern part of Jackson Hole the freshwater Eocene limestone, with its interbedded lava flows, is several times repeated in successive buttes rising above the alluvial floor of the valley. The structure is readily explained by assuming several normal faults with a downthrow on the east, but if the faults exist they are concealed by the alluvial deposits. On the west front of the Hoback Range the Tertiary beds have been faulted down against the Carboniferous. There is abundant circumstantial evidence to indicate that a great fault marks the east base of the Teton Range, cutting off the Paleozoic and Mesozoic

succession to the east, and raising on the west a lofty wall of Archean gneiss. The exposed scarp along this fault is now more than 7,000 feet high (see Figs. 16 and 27) and the total stratigraphic throw can hardly be less than twice that amount.[1]

This deformation evidently took place later than the deposition of the conformable Eocene and Oligocene strata. If disturbances

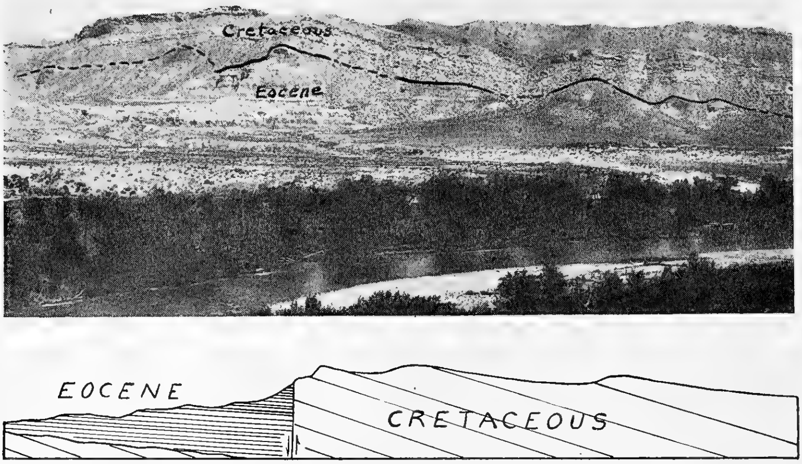

FIG. 9

of this kind had occurred within that depositional interval, they would have caused important unconformities and changes in sedimentation, which are not in evidence. They must also have taken place long before the deposition of the older glacial drift, for the moraines have not been disturbed and on the contrary lie across fault traces along which the scarps have long since disappeared. As will be shown later, allowance must also be made

[1] For a discussion of the question whether this is a recent fault scarp or an old "fault-line scarp" (W. M. Davis, *Bull. Geol. Soc. Am.*, XXIV [1913], 187–216) see below.

here for the making of the high peneplain and the subsequent sculpturing of the modern topography before the last stage of glaciation. Since we cannot yet translate physiographic results into years or even into geologic periods, the portion of the Tertiary and Quaternary periods to be allowed for these events cannot be estimated with any degree of confidence. From the evidence within the district, then, it is justifiable to conclude only that the deformation probably followed the early Oligocene and preceded the Pleistocene.

From a general study of the tectonic disturbances at various times in the earth's history, such as the Appalachian revolution, the

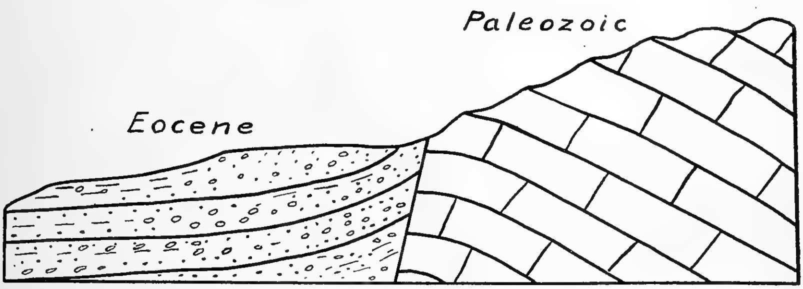

FIG. 10

late Jurassic orogeny of the Pacific, and many others, it seems to be established as a general rule that deformative movements are roughly periodic and that they affect long belts which are continental or even world-wide in their dimensions. It seems fairly safe to assume that the deformative movements evidenced in western Wyoming accorded with this principle, and were manifestations of a much more widespread disturbance, the effects of which should be recognized in adjacent states. It is also self-evident that any such widespread disturbance would cause notable changes in the relief of the land, and would revolutionize the physiographic activities of the region so affected. These changes should now be found in the unconformities or in sudden changes in the character of sediments.

In the Coast Ranges of California, Arnold and others appear to have shown beyond reasonable doubt that sedimentation prevailed

rather generally and without notable interruption from the early Eocene to about the middle of the Miocene as those epochs are now generally correlated, but that near the middle of the Miocene the Tertiary beds were folded strongly so that the late Miocene sediments now rest with marked discordance upon the early Miocene and older strata. In central Washington, Willis, Smith, and

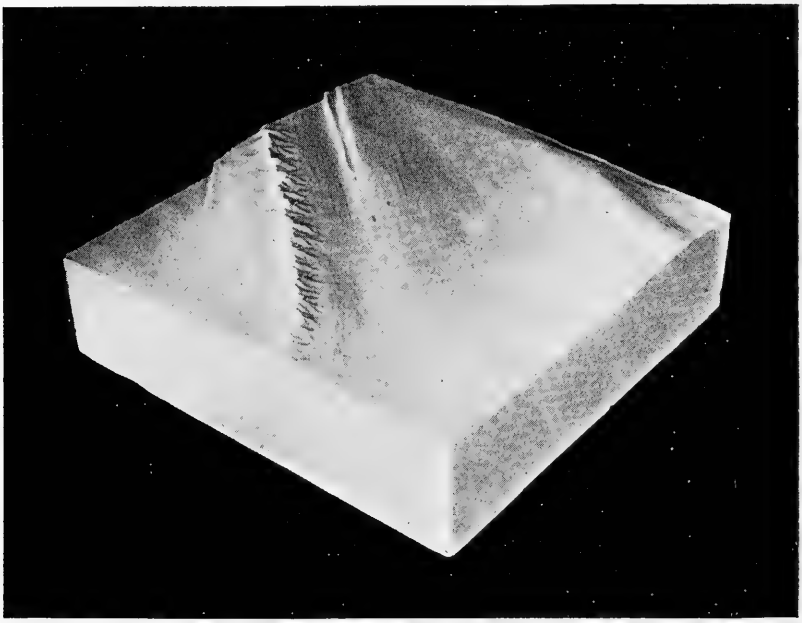

FIG. 11

Calkins have likewise shown that rocks containing early Miocene plants are highly folded and intruded by granites, and that upon their trunkated edges late Miocene lavas and sedimentary beds lie with but little change in their original attitude. In the Grand Canyon region of Arizona the studies of Huntington, Goldthwait, and others lead them to the conclusion that the conspicuous faults of the plateau region were begun long before the late volcanic eruptions, but also long after the gentle folding of the Cretaceous beds. There have also been subsequent movements along the same fractures. A broad study of the Middle Tertiary disturbance has

recently been made by Mr. Hyrum Schneider, and recorded in his manuscript thesis for the M.A. degree at the University of Wisconsin in 1911 (not published). He concludes that the Middle Miocene disturbance affected all of the Cordilleran region, being expressed as folding along the Pacific coast and in Washington, but chiefly by warping and faulting in the Rocky Mountain province.

To summarize these considerations, it may be said then that the sedimentary formations deposited in the early part of the Tertiary period were afterward very gently folded and broken along a few scattered normal faults (see Fig. 11), and that the event took place in the midst of Tertiary time and probably in the middle of the Miocene epoch. Inasmuch as the Tertiary beds almost everywhere dip away from the anticlinal ranges and toward the synclinal basins, it seems evident that the mid-Tertiary deformation merely emphasized the structures produced at the close of the Cretaceous. The faults do not, however, in all cases follow these older structures, for in the Teton Range the original Gros Ventre anticline, which had a northwest-southeast trend, is diagonally transected by several large north-south faults.

[*To be continued*]

THE CANTWELL FORMATION: A CONTINENTAL DEPOSIT OF TERTIARY AGE IN THE ALASKA RANGE[1]

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Introductory statement.—The Cantwell formation comprises a series of conglomerates and finer clastic sediments of important though restricted development within the Alaska Range. These rocks form a belt from 3 to 23 miles in width, which has its beginning near the northeast base of Mount McKinley and extends thence eastward for at least 100 miles to Cathedral Mountain at the head of Susitna River. Its members attain their most typical and maximum expression adjacent to the Nenana Valley, intermediate between those two points.

Historical.—In 1898, during a hasty exploratory trip into the Alaska Range, Elridge noted a "series of conglomerates and coarse sandstones" on the Nenana (then called Cantwell) River, between two tributaries later designated as Jack River and Yanert Fork, and referred briefly to this group of rocks as the Cantwell Conglomerate.[2] Brooks and Prindle in 1902, on a geological reconnaissance of the Mount McKinley region, came across similar rocks at the foot of Muldrow Glacier[3] near Mount McKinley, and following along their northern border traced their extension eastward into the Yanert Fork basin, mapping the area as the

[1] Published with the permission of the Director of the United States Geological Survey. The field work upon which this paper is based was done by F. H. Moffit and the writer, and the latter is indebted to Mr. Moffit for helpful suggestions in its preparation.

[2] George H. Elridge, "A Reconnaissance in the Sushitna Basin and Adjacent Territory, Alaska, in 1898"; *United States Geol. Survey Twentieth Ann. Rept.* (1898–99), Pt. 7, p. 16; also map 3.

[3] They also found a small area 75 miles southwest of that point near Mount Russell of the Alaska Range.

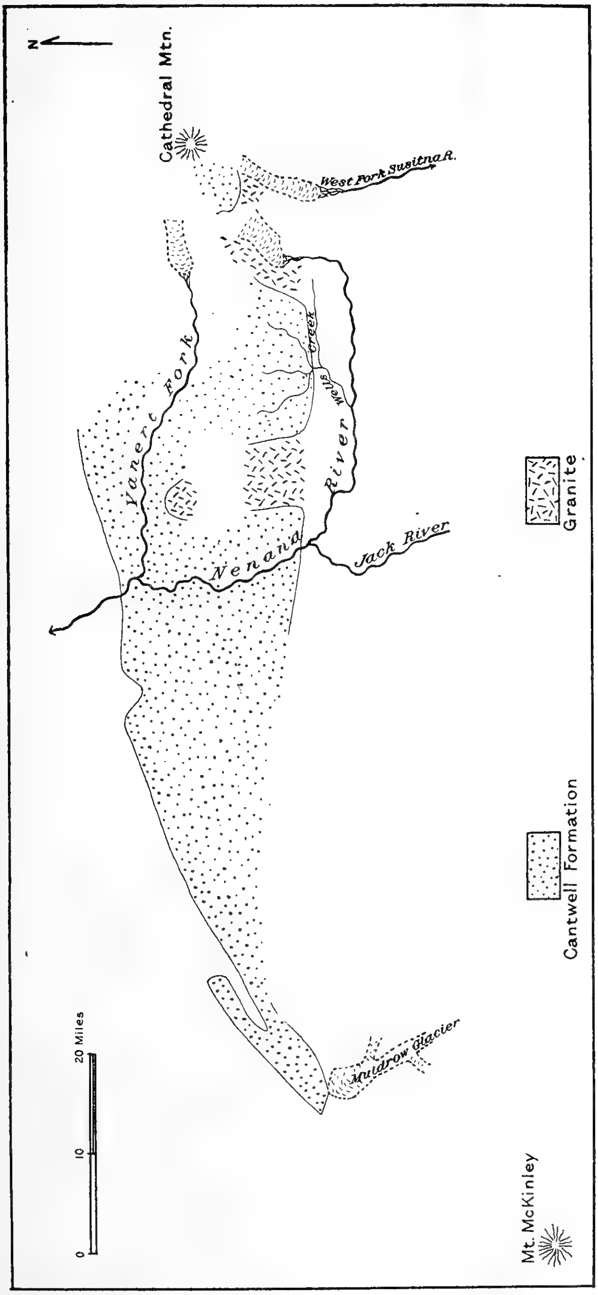

FIG. 1.—Sketch map showing the distribution of the Cantwell formation

Cantwell formation.[1] In 1913 Moffit and Pogue, during a reconnaissance survey of the Broad Pass region, encountered this belt at the mouth of Jack River and studied in some detail its eastward development up to the West Fork Glacier of Susitna River. The present paper deals primarily with this eastward extension of the Cantwell, although it is believed that the conclusions are applicable to the formation in its entirety.

Character.—In its westward extent and along its northern border, the Cantwell formation is characterized by Brooks[2] as follows:

> The Cantwell formation includes a series of heavy conglomerates interbedded with a few shale layers and succeeded by finer conglomerates and red sandstones interbedded with gray and black clay shales. . . . . The basal conglomerate of this formation is made up chiefly of well-rounded white quartz and chert pebbles, the largest of which are 2 inches in diameter. The basal character of the conglomerate is well illustrated in several localities where it rests uncomformably on the older rocks and also contains rounded fragments of them. That part of the formation seen by Elridge contained only the basal beds of the series. On being traced to the north and west these are found to be succeeded by reddish sandstone and gray, drab, and black shales. The sandstones are in some places bright red, but grade from this into a reddish-brown to medium-brown color. The shales are both argillaceous and arenaceous, the latter phase grading into a shaly sandstone. Some of them carry a large amount of carbonaceous matter, and coal seams are interbedded with these rocks, but those seen by the writer appear not to have any commercial importance. . . . .

In the area traversed by the present writer, the Cantwell formation attains its most representative development along the northward course of Nenana River below the mouth of Jack River. The stream here cuts through mountains of conglomerate, which shows in heavy massive beds up to 100 feet and more in thickness, of gentle dip and somber aspect, alternating with subordinate beds of sandstone and argillite. Eastward from the mouth of Jack River the formation changes progressively in character: The conglomerate members become less conspicuous, with increasing prominence of graywacke, sandstone, and carbonaceous shale; while the

[1] Alfred H. Brooks, "The Mount McKinley Region, Alaska," *United States Geol. Survey, Prof. Paper 70* (1911), Plate 9.

[2] *Op. cit.*, p. 78.

rocks no longer remain massive, but show greater and greater schistosity, the finer members passing finally into slates and phyllites, the coarser ones into mica schists.

The massive conglomerate was examined with especial care just east of the mouth of Jack River, where it forms prominent cliffs along the southern face of the Alaska Range. The rock is composed of grains, pebbles, and cobbles (observed up to 7 inches in length) of quartz, quartzite, slate, rhyolite porphyry (?), and perhaps other rocks, set in siliceous cement. Intercalated between these massive beds are thinner members of sandstone, quartzite, and dark-blue carbonaceous argillite. The formation is separated here from the Paleozoic limestone along its southern border by a narrow intrusion of granite, connecting eastward with a large area that intercepts the sediments for a course of several miles.

East of the granite intrusion the Cantwell formation reappears in a more accessible portion of the Alaska Range embracing the headwater region of Wells Creek. This area extends eastward for 13 miles almost to the Nenana Glacier, where it is interrupted by another great granite intrusion. The sediments show the effects of more compression than do the massive beds to the west, and in consequence display considerable schistosity. This Wells Creek area was carefully studied in its western, central, and eastern portions.

In its western portion, a good section was obtained up a small draw from the valley gravels to the mountain top. The lowermost 1,500 feet consists predominantly of massive to somewhat schistose graywackes of dark-gray color, composed of subangular to rounded pieces of quartz and black slate set in siliceous cement. This rock carries intercalated beds of black carbonaceous shales with abundant impressions of leaves and of siliceous conglomerate, in part somewhat schistose. The remaining 600 feet to the mountain top is composed of prominent, heavy beds of a severely mashed phase of the conglomerate, light greenish to yellowish in color, in places so schistose that the original texture is nearly obliterated. The average rock, however, shows the coarser components as elongated lenses, resembling so-called "stretched-pebbles," enwrapped by the finer constituents, which have been rendered in part

micaceous. The beds dip northward at angles of 35° to 60°, indicating that the folding is less gentle than to the west.

In its central portion the Wells Creek development of the Cantwell was again examined in detail. The adjoined section of the lowermost portion in contact with underlying granite indicates the rapid variation in conditions that gave rise to this series.

SECTION OF LOWERMOST PORTION OF CANTWELL FORMATION NEAR WELLS CREEK

Schistose conglomerate	15 feet
Silicified tuff[1]	15 "
Alternations of schistose conglomerate and schistose graywacke, carrying beds of shattered black slate and lens-like intercalations of tuff	50 "
Dense, light-colored tuff, very hard	5 "
Fairly schistose conglomerate, grading into schistose graywacke, which in turn grades into schistose conglomerate; the whole carrying pinching and swelling lenticular beds of tuff	15 "
Graywacke, with lenses of quartzite and of shattered carbonaceous slate	30 "
Fairly massive siliceous conglomerate	2 "
Graywacke	3 "
Carbonaceous slate, badly shattered and slickensided	8 "
Graywacke, slightly schistose	15 "
Crushed slate and schistose graywacke	5 "
Graywacke, fairly massive	5 "
Covered, mostly altered graywacke	20 "
Schistose and altered graywacke, injected with quartz stringers and lenses	10 "
Granite, exposed	30 "
Glacial gravels ... to valley bottom.	
	228 feet

For a few hundred feet above this section the rocks are not exposed; then there appears a narrow bed of shale, bordered by a large barren quartz vein, which shows beneath a heavy somewhat contorted bed of comparatively massive conglomerate composed predominantly of white quartz pebbles averaging three-

[1] As shown by the microscope, this rock is composed of small pieces of quartz, with some orthoclase and a very little plagioclase—all distinctly fragmental—set in a dense, microcrystalline ground.

fourths of an inch in diameter. From this exposure to the mountain top, a vertical distance of about 1,100 feet, the rocks are dominantly black shale with plant impressions, alternating with beds of massive graywacke, in some places approximating a sandstone. Near the eastern base of this mountain the carbonaceous, plant-bearing shales were observed to carry 2- to $2\frac{1}{2}$-inch seams of impure lignitic coal. For $2\frac{1}{2}$ miles north of this point mashed conglomerate alternates with dark-blue to black slate (resulting from the metamorphism of the carbonaceous shales), some phases showing a crinkly surface of satin-like luster. Many of the rocks here might properly be termed schists, and the formation as a whole presents a black aspect due to the widespread dissemination of carbonaceous material.

A traverse was carried up a stream farther eastward in the Wells Creek area. Near the southern border of the formation, this stream exposes closely folded beds of slate, graywacke, and conglomerate, all severely mashed, the conglomerate almost beyond recognition. The slate is also badly crushed and slickensides are prominently developed. For two miles above these outcrops the formation seems about equally divided between slate, graywacke, and conglomerate, the last two having suffered most from the regional metamorphism, although much of the slate has a crinkled surface, suggesting the intensity of the compression to which the rocks were subjected. Throughout this extent the rocks are steeply dipping and apparently vary from 80° N. to 80° S. In places are crinklings and plications suggestive of sharp folding, and many specimens of a peculiar, plicated phase of mashed conglomerate were observed in talus fans. The formation as a whole is dominantly dark-grayish to black in color.

East of the Wells Creek basin, the Cantwell is replaced by a large intrusion of granitic character, forming a rugged mountain range surrounding the Nenana Glacier. A few miles northeast of the glacier, at the foot of Cathedral Mountain, the Cantwell apparently appears again; this area was not directly accessible, though observable with field glass and through the agency of glacially transported specimens. The sediments here consist of dark-colored slates and mica schists, the latter representing the coarser clastics

severely mashed. A typical example of the mica schist is dark-reddish to bluish-gray in color, weathering to an iron-stained surface glistening with mica scales along its planes of foliation and presenting on transverse break thin quartzose lenses, deformed survivors of original pebbles. It is only after having traced the Cantwell through its progressive stages of metamorphism from west to east that an observer would recognize this type as the mashed equivalent of the massive conglomerate so extensively developed near the mouth of Jack River, 35 miles to the west.

Associated igneous rocks.—In the western extension of the formation lava flows of andesitic, rhyolitic, and basaltic character are interbedded with the conglomerate, and these intercalations are especially prevalent toward the Nenana Valley, in some localities equaling the sediments in thickness.[1] West of the Nenana the Cantwell is cut by dikes of diabase and stocks of granite. Reference has already been made to the granite batholiths that invade the Cantwell sediments east of the Nenana Canyon. Adjacent to these areas the sediments are traversed by abundant dikes of rhyolite porphyry and diorite porphyry, the former especially conspicuous near the head of Wells Creek.

Structure.—The structure of the Cantwell formation has been suggested in the preceding paragraphs. Its low northward dip and unmetamorphosed character on the west, with increasingly steeper dips and development of schistose textures along its eastward course, point to folding, gentle to the west, more and more intense to the east. The trend of the formation, and of the axis of folding, as indicated by numerous strike measurements, as well as by the course of the southern boundary of the formation, is about N. 80° E.

The inclination of the massive conglomerate beds where the Nenana River turns northward and cuts through them indicates that the southern face of the mountains here represents the beveled upturned edge of a gentle synclinal fold. According to Brooks[2] this type of broad open folding holds for the westward extension of the formation. At the head of Wells Creek the Cantwell is structurally more complicated than in the area just noted. It

[1] Brooks, *op. cit.*, p. 79.

[2] *Op. cit.*, p. 79; also section CD on geologic map, Plate 9.

here shows steep, though variable, dips, both to the north and to the south; and transverse to the trend of the sequence displays considerable variation in the degree of dynamic metamorphism, the rocks ranging from fairly massive to schistose. These observations indicate that the rocks have been subjected to considerable compression, and it is believed that the gentle folds on the west pass eastward into more intense and complicated folds, not isoclinal, but with axial planes variously inclined on an east-west axis.

Although no faults were directly observed in the portions of the Cantwell examined, faulting was undoubtedly a resultant of the compressive forces to which the terrane has been subjected. It would appear from the increasing pressure effects to the east that the formation as a whole had suffered a differential degree of compression. Yet the possibility must not be overlooked that in the massive areas extensive faulting may have relieved the stresses that in the absence of these places of yielding would have produced close folding and schistosity there also. In the westward part of the formation (west of the Nenana), where the rocks are massive and folding gently, Brooks[1] observed a number of faults consequent upon the folding.[2]

Contact relations.—Along the northern border, according to Brooks, the Cantwell rests unconformably upon Paleozoic rocks. Along the eastern portion of the southern border, the relation to the bordering Paleozoic sediments is obscured by granite intrusions and glacial gravels, but the contact is believed to be characterized by an unconformity, although there is some evidence that the plane of separation is one of fault movement. The coming to place of the extensive granite batholiths seems to have had little structural or mineralogic effect upon the invaded sediments.

Thickness.—The thickness of the Cantwell formation was obtained with greatest accuracy in the section east of the mouth of

[1] *Op. cit.*, pp. 79–80.

[2] Brooks (*op. cit.*, p. 79) suggests that the Cantwell is not so sharply folded as the adjacent formation, "because the massive beds were able to resist the movement, some of which has been taken up by lines of shearing that have followed the shale beds and the lines of parting between the shales and the massive beds."

Jack River, where calculations based upon aneroid and triangulation measurements show a vertical development of about 2,700 feet. In the headwater region of the Toklat basin about 25 miles to the west, Brooks[1] found that the thickness there exposed is at least 2,000 feet. These figures do not represent the total thickness of sediments as deposited, because the top of the formation is a surface of erosion, from which an uppermost portion of unknown thickness has been removed.

Age and correlation.—The Cantwell formation is assigned to the Tertiary on the basis of plant remains found in some abundance in the Wells Creek area. The following forms were identified by Drs. F. H. Knowlton and Arthur Hollick, who report that the material is of Tertiary age:

Taxodium tinajorum Heer.
Taxodium dubium (Sternberg) Heer?
Sequoia langsdorfii (Brongniart) Heer?
Populus arctica Heer?
Daphnogene Kanii Heer.
Aspidium Heerii Ettingshausen?
Ginkgo adiantoides (Unger) Heer?

In the report on the Mount McKinley region[2] the Cantwell was provisionally referred to the Carboniferous, although its lithologic resemblance to several occurrences of the Kenai formation (Eocene) was noted and the possibilities of younger age fully recognized. The extent and relation of the plant-bearing area surrounding upper Wells Creek and the fact that a locality within the western portion of the Cantwell has also furnished Tertiary plant remains showing identical species[3] would seem to fairly establish the present age assignment.[4]

[1] *Op. cit.*, p. 81. [2] *United States Geol. Survey, Prof. Paper 70*, 1912.

[3] Brooks, *op. cit.*, p. 82. This occurrence was interpreted as an infaulted block and the fossil evidence was not regarded as applicable to the surrounding formation.

[4] The possibility must be recognized that future detailed work may result in a subdivision of the Cantwell, as in its western portion in particular the members in which conglomerate beds dominate are succeeded by a series of shales and sandstones. It seems likely, however, that these upper members are merely an expression of normal, changing topographic conditions in the region of supply, rather than the result of a distinctly later epoch of sedimentation. In its eastern portion, where observed by the present writer, the formation appears definitely a unit.

The Cantwell as indicated by its plant remains and character is of Eocene age and corresponds in whole or in part to the Kenai formation, widely distributed in isolated areas through Alaska. This correspondence, however, should be regarded as a general one, without implication of exact time coincidence or precise analogy of formations; for the Cantwell formation and the Kenai, being throughout dominantly continental deposits, cannot be expected to display the degree of equivalence normal to marine deposits.

Origin.[1]—The character of the sediments composing the Cantwell formation signifies fluviatile continental deposition. The extensive development, number, and thickness of the conglomerate beds and the heterogeneous character of this rock indicate a source of supply under violent degradation, a short transport, and distribution by rushing waters competent to spread material of such coarseness. The presence, moreover, between the conglomerate members of carbonaceous mud rocks, carrying well-preserved plant remains in abundance and variety, implies areas of quiet deposition, such as swamps and small lakes, which alternately superseded, and in turn were transgressed by, zones of depositional activity. These conditions are afforded by voluminous streams emerging from a mountainous region on to a piedmont slope, intermontane valley, or shallow sea.

It is difficult with present information to decide whether the Cantwell was laid down as an interior continental deposit or upon the inland margin of a delta built out from a coast range of mountains. The apparent absence of deposits that might be attributed to the seaward portion of such a delta, the predominance of conglomerate over finer clastics, and topographic considerations incline the writer to favor the first explanation; although on this basis, for the preservation of the sediments from subsequent complete erosion, it must be assumed that downward warping in front of the mountain axis succeeded their deposition. That a compressive

[1] The interpretation of the origin of the Cantwell formation has been facilitated by a study of the contributions of Barrell on sedimentation, especially those appearing in *Jour. Geol.*, XIV (1906), 316–56, 430–57, 524–68; XVI (1908), 159–90, 255–95, 363–84; *Bull. Geol. Soc. Amer.*, XX (1909), 620; XXIII (1912), 377–46; and *Amer. Jour. Sci.*, XXXVI (1913), 429–72; XXXVII (1914), 87–109, 225–53. See also Mansfield, *Jour. Geol.*, XV (1907), 550–55.

earth movement followed deposition is clearly evidenced by the present folded character of the sediments.

The freshness and variety of the components of the conglomerate and finer sediments point to derivation from a region of rugged and youthful[1] topography under conditions of intense mechanical disintegration characteristic of a climate in which frost action played an important part. The presence of carbon and plant remains in the finer members of the formation indicates deposition under conditions unfavorable to oxidation, while the character of the plants themselves implies a temperate climate in the area of deposition, markedly milder than that of the region today, and in contrast also to the more rigorous conditions that prevailed in the high mountains from which the sediments were derived. The thickness and extent of the conglomerate members, moreover, seem to indicate a more vigorous precipitation than at present, capable of handling the abundant waste of the mountains. That this upland was the seat of active glaciation and consequently contributed glacial materials to the streams sweeping the Cantwell sediments downward is regarded as probable, though no observations can be adduced in support of this belief.

The presence of red sandstones toward the top, as noted by Brooks in the western portion of the formation, is not believed to have climatic significance;[2] for organic colors are conspicuous throughout that portion of the formation also. These members are rather suspected of expressing a variant in the lithology, or a change in the topography, of the source of supply.

The situation of the Tertiary land mass in reference to its basin of deposition cannot be given without reservation, but the present delineation of the Cantwell formation and the fact that its coarser members contain fragments of slate similar to that of older formations to the south and dissimilar to sediments exposed to the north suggest a position south of the present area of Tertiary rocks.

[1] See Willis, *Jour. Geol.*, I (1893), 478.

[2] Barrell (*Jour. Geol.*, XVI [1908], 293) observes that ". . . . red shales or sandstones, as distinct from red mud and sand, may originate under intermittently rainy, subarid, or arid climate without any close relation to temperature and typically as pluvial or fluvial deposits upon the land. . . . ."

THE PHYTOSAURIA OF THE TRIAS

MAURICE G. MEHL
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Some time ago the writer gave a brief notice of a new genus of phytosaurs of which *Angistorhinus grandis* Mehl was the type.[1] It is the purpose of this paper primarily to give a fuller description of this form and of another specimen mentioned in the above paper which further study has shown to be a new species of the same genus.

Angistorhinus grandis MEHL

The general characteristics of the specimen upon which this form is based were set forth in the previous paper and deserve but brief mention here. The skull is elongate with the rostrum produced into a long, slender, depressed snout and the nares elevated on a prominence at its posterior end. It is among the largest of the phytosaurian skulls, with a total length of about 977 mm. The squamosals extend a considerable distance beyond the posterior border of the quadrates and are produced downward into stout, hooklike processes. The supratemporal vacuities are closed posteriorly by a well-developed parieto-squamosal arcade that lies in the plane of the roof of the cranium. A marked depression is seen on the dorsal surface surrounded by the orbits and supratemporal vacuities. The irregular pitting of the surface is confined almost entirely to the lateral and posterior sides of the prominence upon which the nares are situated, and the flat dorsal surface of the cranium back of this, and in front of the supratemporal vacuities. In a lateral view the skull resembles that of *Mystriosuchus* Fraas[2] more than any of the other phytosaurs.

[1] *Jour. Geol.*, XXI (1913), 186.

[2] *Die Schwabischer Trias-Saurier nach dem Material der Kgl. Naturalien-Sammlung in Stuttgart zusammengestellt* (1896), p. 16, Pl. V.

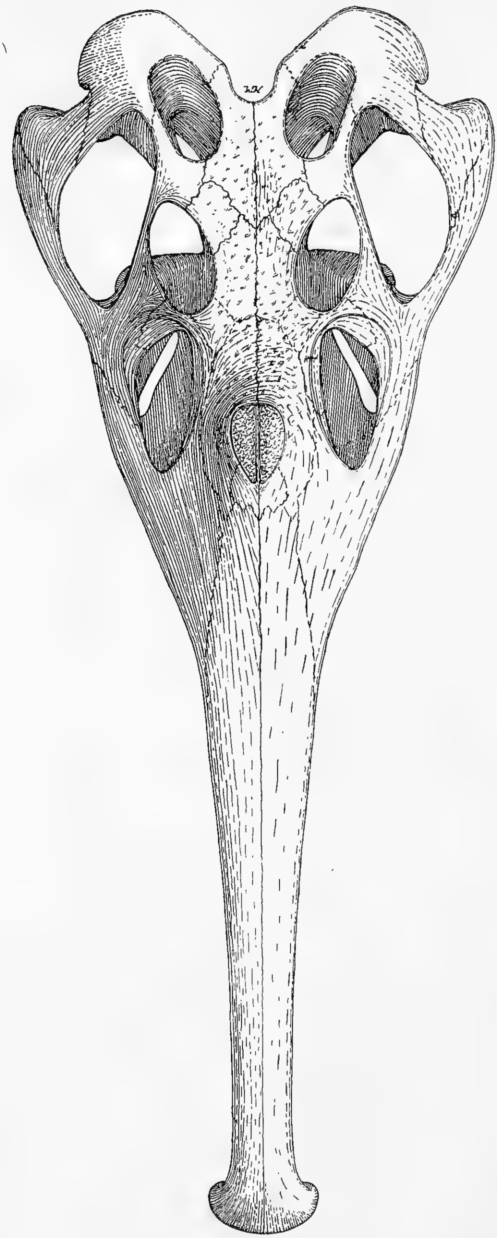

FIG. 1.—*Angistorhinus grandis*, dorsal view of skull, about one-sixth natural size

OPENINGS OF THE DORSAL SURFACE OF THE SKULL

The *external nares* are situated on a prominence at the beginning of the rostrum and rise fully 18 mm. above the level of the general surface of the cranial roof. The thin, vertical, median septum separating the nares does not rise to the same height as that of the posterior and lateral borders and thus the nares are given the appearance of a single opening with rounded posterior and acute anterior border. In length they are about 62 mm. and the greatest width of the seemingly common opening is 40 mm. The anterior border is about even with the anterior borders of the antorbital vacuities and 590 mm. from the tip of the rostrum. The region about the anterior border of the nares is somewhat abraded, and the sutures are not distinguishable. No septomaxillae have been distinguished, but as they are present in another specimen to be described later they probably have about the same position here and form the anterior border of the nares.

The *antorbital vacuities* are large, oval in outline, with rather acute anterior and posterior extremities. Their greater diameter is fully 130 mm. and the lesser diameter about 55 mm. Their size is accentuated by an abrupt depression or excavation of the bone along their posterior and upper borders.

The large *orbits* are somewhat elongate antero-posteriorly, about 89 mm. long and 55 mm. wide. Their planes are directed outward and a little less upward, perhaps. The interorbital width is about 68 mm. Only a narrow bar, not over 9 mm. in width, separates the orbit from the lateral temporal fenestrae.

The sides of the *lateral temporal fenestrae* form rough parallelograms with unequal diagonals. The greater diagonal is 171 mm. and the lesser, which is almost perpendicular to the plane of the palate, is about 113 mm. The greatest width of these openings is about 87 mm. There is an excavation for a short distance about their lower anterior borders similar to that of the upper posterior border of the antorbital vacuities.

The *supratemporal fenestrae* are oval in outline with an antero-posterior diameter of 84 mm. and a width of 46 mm. Their plane is directed upward. They are peculiar in that they are closed behind by a well-developed parieto-squamosal arcade that lies in

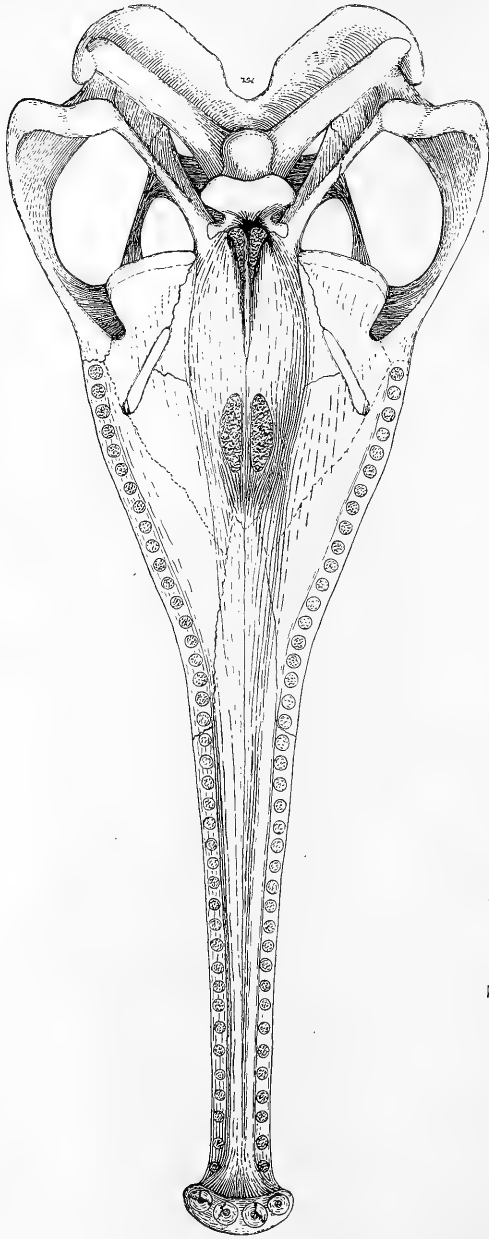

Fig. 2.—*Angistorhinus grandis*, ventral view of skull, about one-sixth natural size

the plane of the dorsal surface of the cranium. This will be discussed more fully in connection with the posterior aspect. No *parietal foramen* is present.

THE SEPARATE BONES OF THE DORSAL SURFACE

The *parietals* are united along the median line for a distance of about 100 mm. Posteriorly they are separated for a short distance by a deep, rounded notch. In spite of this deep incision, however, the occipital condyle is not visible in a dorsal view. Their anterior ends form an acute angle that separates the frontals for a short distance. The greatest width of the parietals, a little in front of the anterior border of the supratemporal fenestrae, is about 42 mm. for each element. The parieto-squamosal suture is not clearly distinguishable; apparently these elements meet on the inner posterior borders of the supratemporal fenestrae.

The *squamosals* form the posterior and half of the lateral borders of the supratemporal fenestrae and the upper posterior borders of the lateral temporal fenestrae. They are bent abruptly downward at a point on the antero-posterior line of the bar separating the supratemporal and lateral temporal fenestrae, the downward extension being produced into a hooklike process similar to that of *Mystriosuchus* but with greater development. These hooklike angles extend below the dorsal surface of the skull a distance of about 104 mm. This part of the squamosal and the lower posterior surface of the quadrate as well was weathered away, but a perfect natural mold in the matrix made possible a very accurate restoration.

The *postorbitals* form a part of the antero-lateral borders of the supratemporal fenestrae and the postero-lateral borders of the orbits. They extend down and forward to meet the upper posterior process of the jugal, and the two form narrow bars that separate the orbits and the lateral temporal fenestrae. The inner boundaries of the postorbitals are formed by the postfrontals and the parietals which meet near the middle of the postorbitals' length.

The *postfrontals* are rather small, approximately quadrangular elements that form the posterior half of the upper borders of the

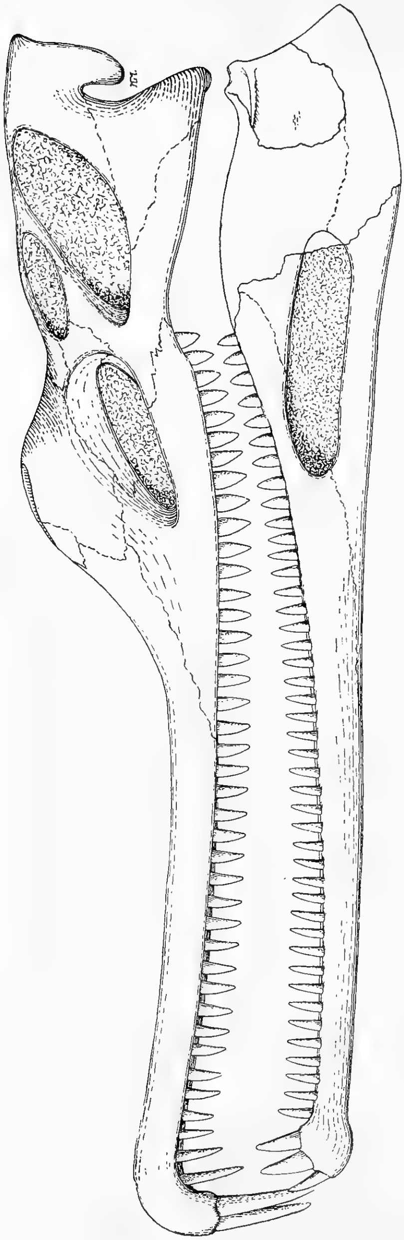

FIG. 3.—*Angistorhinus grandis*, lateral view of skull and mandible, about one-sixth natural size

orbits. They are bounded on the inner side from front to back by the frontals, the parietals, and the squamosals. The anterior sutures of the frontals are not evident.

The *frontals* appear to extend forward to within 68 mm. of the nares. This would give them a length along their median union of about 68 mm. They form the upper borders of the orbits and join the prefrontals and nasals in front. Their anterior margin seems to form a slight concavity into which the posterior ends of the nasals extend.

The *prefrontal* sutures are not certain; these elements seem to be small, approximately quadrangular in outline, and form the upper anterior borders of the orbits.

The *lachrymals*[1] unite with the maxillae to exclude the nasals from taking part in forming the borders of the antorbital vacuities. Their downward extent and union with the maxillae are uncertain.

The *jugal* forms the lower postero-lateral border of the skull. At its anterior end the jugal sends forward a splinter-like process that forms the lower posterior border of the antorbital vacuity. It also sends back from the anterior upper side a process that unites with the lachrymal in front and the postorbital behind, and thus forms the lower anterior border of the orbit and the anterior half of the lower temporal fenestrae boundaries.

The *quadratojugal* connects broadly with the jugal below. Its suture with the squamosals is not distinguishable. The sutures between the nasals, the septomaxillae, the maxillae, and the premaxillae cannot be determined with certainty. In all probability the relations are much the same as in another form to be described later (see Fig. 4).

The *premaxillae* are produced into a long, slender, subcylindrical snout much like that of *Mystriosuchus*. Viewed dorsally it tapers gradually from the cranium proper to a point near the extremity, from a width of 94 mm. at a distance of 140 mm. in front of the antorbital vacuity, to 47 mm. Near the tip the rostrum expands

[1] E. Gaupp has attempted to show that the so-called "prefrontal" of reptiles is homologus with the mammalian lachrymal and has suggested the name "adlachrymal" for the element previously called the lachrymal in reptiles (*Anatom. Anz.*, XXXVI, 1910). According to Gregory (1913), however, this homology is in nowise proven. In this paper the writer adheres to the old nomenclature.

gradually, attaining finally a width of 84 mm. The anterior extremity is bent abruptly downward and reaches a distance of 39 mm. below the plane of the ventral surface.

THE PALATE

Within the line of the alveoli on the ventral surface is a prominent rounded ridge on either side, as is noted in *Palaeorhinus bransini* Williston[1] and *Mystriosuchus*. It crowds close along the alveoli throughout the entire length of the premaxillae and gradually flattens

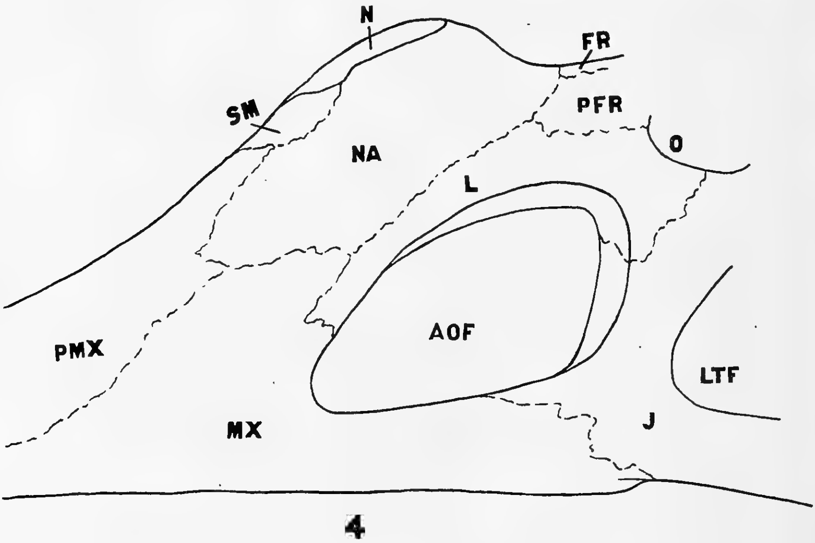

FIG. 4.—*Angistorhinus gracilis*, lateral view showing apparent arrangement of bones about the narial region, one-third natural size.

out on the maxillae. The teeth have all dropped from the alveoli except in the down-curved portion of the rostrum. Here four large ones are to be seen broken off close up to the premaxillae, two on each side. They are all round in section and measure about 15 mm. in diameter. If one may judge from a similar specimen in which these terminal teeth are preserved, they reached a length of from 80 mm. to 90 mm. or more in this form. In the premaxillae there are 23 alveoli on each side. These, with the exception of the two anterior ones described above, are approximately of one size, aver-

[1] J. H. Lees, "The Skull of Palaeorhinus," *Jour. Geol.*, XV (1907), 124.

aging about 11 mm. in diameter. The third tooth of the premaxillae series crowds close upon the down-curve portion of the rostrum. Back of this they are quite regularly spaced, averaging about 8 mm. between adjacent alveoli. A break across the rostrum near the maxilla-premaxilla union exposed an unerupted tooth very similar to that shown in Fig. 9.

The *maxillae* have their greatest extent along the lateral margins of the skull, where they reach a length of about 340 mm. They connect with the jugals near the middle of the antorbital vacuities and form the lower anterior borders of these openings. Above they

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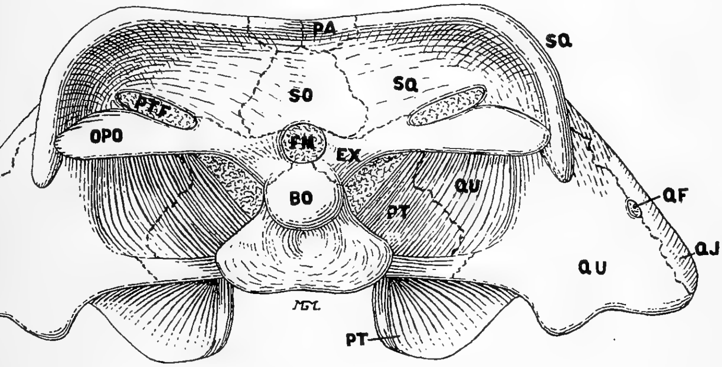

Fig. 5.—*Angistorhinus grandis*, posterior view of skull, two-sevenths natural size

connect slightly with the long anterior projections of the lachrymals and more broadly with the nasals. They contain about 19 alveoli each, of about the same diameter as those of the premaxillae. The spacing is much the same with the exception that in the posterior ones there is a slight diminution of the interalveolar space. Although little can be said with certainty concerning the crowns of the premaxilla-maxilla teeth, they probably ranged from those with nearly round sections, in the anterior part of the rostrum, to those that are laterally compressed with sharp, serrate anterior and posterior cutting edges. A great many of these sorts are found associated with the phytosaur remains and some of the laterally compressed ones were found in a fragment of a skull that very closely resembles the form here described. These teeth will be more fully discussed farther on.

The *palate* is somewhat crushed and distorted and most of the sutures are indistinguishable. Still a fairly good idea of its configuration and the general relations of the bones can be gained. It is quite highly arched along the median line and especially so in the region of the nares. The low, rounded arch along the median line of the premaxilla and maxilla, formed by the ridges parallel to the alveoli, increases gradually in height and width posteriorly, the increase being more rapid as the nares are approached, till at the posterior margin of the nares the arch is about 82 mm. wide at the general level of the palate surface and rises to a height of about 42 mm. from that plane. Behind this point the arch attains a still

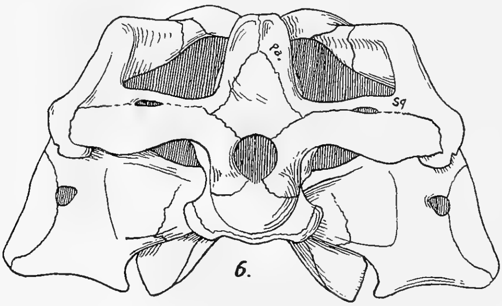

FIG. 6.—*Mystriosuchus planirostris*, posterior view of skull, two-sevenths natural size. Outlines and sutures after J. H. McGregor.

greater width before it again rounds in posteriorly. The internal nares are a little back of the external nares, 10 mm. perhaps. The forward ends of these openings are still covered by the matrix. Their exact extent in that direction cannot be told, but they are probably about 48 mm. long. The width between their lateral borders is 34 mm. The median partition is slightly wider on the palate surface than it is between the external nares. A line connecting the anterior borders of the *post-palatine foramina* crosses the internal nares a little in front of the posterior border. From this line the foramina extend back and inward a distance of 60 mm. They are about 8 mm. wide, are slightly curved, concave inward, and have rounded anterior and posterior borders. Their anterior extremities are 190 mm. apart and their posterior ends some 120 mm. In size they are similar to those of *Mesorhinus fraasi*

Jaekel,[1] a little larger, however, and more slitlike. In both of these forms the post-palatine foramina are considerably larger than in most of the other phytosaurs. The inter-pterygoid vacuity is notable in this form for its antero-posterior extent, being about 103 mm. long. It gradually diminishes in width from 35 mm., its greatest lateral extent, near its posterior boundary, to a very acute angle about 30 mm. back of the posterior border of the nares.

THE SEPARATE BONES OF THE PALATE

The posterior extension of the *premaxilla* is indeterminate; apparently it lacks quite a little of reaching the anterior border of the nares.

The *vomers* seem to have little lateral extent. The exact condition cannot be seen, however, as none of the boundaries is distinguishable. In all probability they unite broadly in front with the premaxillae and form most of the anterior border of the nares as well as their median borders. They probably do not extend as far back as the posterior end of the nares, certainly they cannot extend any considerable distance beyond.

Apparently the maxillae do not enter the nares at all. This is a condition suggested by Lees in *Palaeorhinus bransoni* (*op. cit.*, Fig. 2). If they do take part in the boundaries of these openings it is very slight and along the antero-lateral borders. The pterygo-palatine sutures are also more or less hypothetical.

The *palatines* reach far forward, forming most or all of the lateral borders of the nares. They seem to join the pterygoids posteriorly about the middle of the posterior palatine foramina. At the anterior ends of these openings they connect very slightly with the transverse (ectopterygoid) bones, if at all. On the general plane of the palate their greatest width is about 52 mm. along a line connecting the anterior ends of the posterior palatine foramina. At this width they are bent almost vertically upward in an antero-posterior direction.

The *pterygoid bones* form the walls and roof of most of the spacious palate arch back of the nares. Only a small triangular

[1] "Über einen neuen Belodonten aus dem Buntsandstein von Bernburg," *Sitzungsberichten der Gesellschaft naturforschender Freunde*, No. 5, Jahrgang, 1910, p. 209, Fig. 5.

portion of them is to be seen on the general plane of the palate surface. They seem to form the posterior half of the inner boundaries of the posterior palatine openings. Near the anterior limit of the basisphenoid they bend abruptly upward from the roof of the broadly rounded palate arch and extend above the parasphenoid. Anteriorly they approach each other till they embrace the para- and presphenoid and, still farther forward, close together below them. They meet the transverse bones broadly at the posterior ends of the posterior palatine foramina, the suture being directed backward from this point and a little inward. Posteriorly they close with ample and firm contact on the basi-pterygoid process. From this union the pterygoids send back and outward narrow processes which meet like processes of the quadrates about midway between the quadrates proper and the pterygo-sphenoid union. Their relations with the various elements seen in a posterior view will be discussed below.

The *transverse bones*, seen from below, present an approximately triangular outline. They form the lateral borders of the posterior palatine foramina and the anterior borders of the post-temporal vacuities. The broad pterygo-transverse sutures run back and inward from the posterior ends of the posterior palatine openings toward the basi-pterygoid process. Along the outer anterior side the transverse bones unite with the maxillae for a distance of about 40 mm. and at a somewhat less distance with the jugals. The postero-lateral corners are bent downward along a line running from their inner posterior angles obliquely outward from the direction of the posterior palatine openings and extend some 14 mm. below the general ventral plane of the skull.

The ventral surface of the *parasphenoid* is exposed the entire length of the inter-pterygoid vacuity and both the para- and the presphenoid are free from the matrix for some distance along one side. Their union is clearly indicated by a shallow longitudinal groove. At their union with the basisphenoid they form a thin, vertical plate which gradually thickens anteriorly very slightly and becomes rounded below. They are visible from below for a distance of 95 mm. Anteriorly at this distance, about 42 mm. back of the posterior border of the nares, the pterygoids meet below these

elements and hence their full anterior extent cannot be determined. At their point of disappearance, however, they still maintain their full thickness and must extend farther forward for some distance. Viewed from below, the postero-lateral processes of the sphenoid, the greater wings, form a high, transverse, rather sharp ridge slightly depressed at the middle and somewhat thickened at the extremities. This is about 76 mm. long and at the middle extends

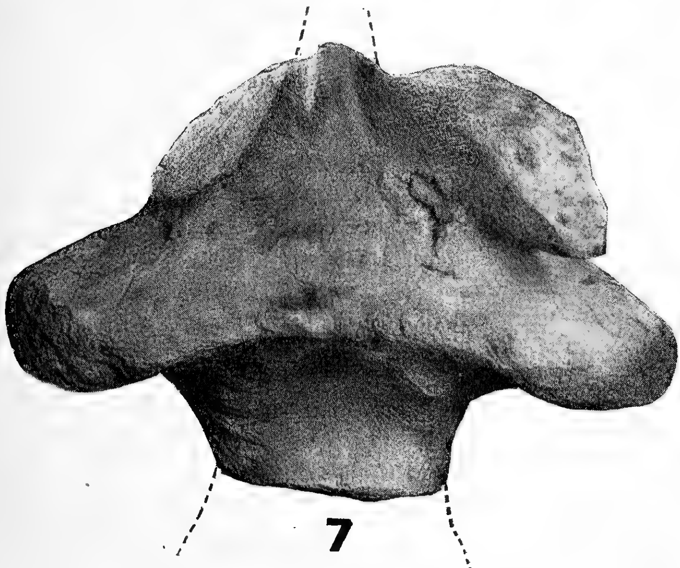

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FIG. 7.—*Angistorhinus grandis?* Cojoined basioccipital and basisphenoid, natural size.

down 44 mm. below the lower edge of the presphenoid. The anterior face of the pterygoid processes of the sphenoid reaches forward about 35 mm. from this ridge and extends down about even with it. These processes are quite stout and are separated from the posterior wings by a narrow but deep channel. The posterior part of the basisphenoid has been considerably abraded in this specimen and a portion of the occipital condyle is missing. Another specimen is present in the collection, however, a co-ossified basioccipital and basisphenoid (see Fig. 7) that resembles this form very closely, and this has aided materially in the interpretation of the specimen at hand. On the posterior side of the transverse plate formed by the

greater wings of the sphenoid is a circular excavation 26 mm. in diameter. Although shallow the depth of the excavation is accentuated by a prominent curved ridge on either side springing from the base of the occipital condyle.

THE POSTERIOR ASPECT

In a posterior view *Angistorhinus* resembles in the general form *Lophoprosopus*[1] (*Phytosaurus*) *kapffi* H. von Meyer[2] or *Mesorhinus Fraasi*. It is depressed in appearance as the skull is over twice as wide as high. From the lowest point of the quadrates to the plane of the dorsal surface it measures about 166 mm. Between the lower, lateral angles of the quadrates the width is about 386 mm. The post-temporal fenestrae are exceptionally large, perhaps larger than in any other form. Those in *Lophoprosopus kapffi* (see reference above) are about 32 mm., while in *A. grandis* they measure nearly 40 mm. The width is about 11 mm. From the outer extremities of these openings, which are about 180 mm. apart, they are directed inward and slightly down. The foramen magnum is smaller than in most of the phytosaurs, about 20 mm. in diameter. The quadrate foramina lie between the quadratojugal-quadrate union, about 46 mm. above the lower face of the quadrates. They are oval in shape, the vertical diameter being 23 mm., the lesser diameter about 11 mm. The outer border of these foramina is formed by the quadratojugals, the inner border apparently by the quadrate alone. One of the most remarkable features seen in the skull of *A. grandis* is the difference between this form and most of the other phytosaurian genera in the upper posterior border of the skull. In *Lophoprosopus*, *Mystriosuchus*, and other forms, perhaps, the parieto-squamosal arcade, bounding the upper temporal vacuity posteriorly, is considerably reduced and depressed. Quoting McGregor on this point in *Mystriosuchus planirostris* Fraas:[3]

A casual observation of the dorsal surface of the skull might lead the observer to think that the supratemporal fenestra was incomplete posteriorly,

[1] The substitution of *Lophoprosopus* for *Phytosaurus* is explained in the appended list of genera and species.

[2] McGregor, *Memoirs of the Amer. Mus. Nat. Hist.*, Vol. IX, Part II (1896), p. 45, Fig. 4.

[3] *Ibid.*, p. 46.

though its outer, anterior, and inner margins are normally represented by the highly sculptured squamosal, postorbital, and parietal. The median posterior portion of the parietal seems to end suddenly over the occiput, and the parieto-squamosal arcade appears to be lacking. In fact, Marsh has stated (1896) that "the skull of *Belodon* [*Phytosaurus*] shows that the supratemporal openings, characteristic of the true crocodillians are wanting."

Again, in characterizing the Phytosauria, McGregor says:[1] "Post-temporal (parieto-squamosal) arcade greatly depressed, reduced, and covered by muscles." In *Angistorhinus grandis*, however, this is not the condition. A comparison of the posterior view of *A. grandis* with that of *Mystriosuchus planirostrus* (Figs. 5 and 6) will serve well to show the differences. In the former the post-temporal arcade is well developed and borders the supratemporal vacuities in the same plane that extends over the roof of the cranium. The posterior extension of this parieto-squamosal arcade forms an overhanging shelf that in a dorsal view hides all the bones of the posterior side of the skull. The *parietals* seem to have a comparatively small lateral and downward expansion on the posterior surface and but slight contact with the supraoccipital. The squamosals, on the contrary, send down broad, platelike extensions from the posterior borders of the supratemporal vacuities that connect broadly with the supraoccipital at their inner edges. They form the upper and outer borders of the post-temporal vacuities and connect broadly with the upper inner ends of the opisthotics. As the specimen is now prepared, a large fenestra pierces this element on either side of the skull just over the post-temporal vacuity. This opening is, however, undoubtedly due to the abraded condition of the bone as the matrix beneath indicated an uninterrupted bone surface. The lateral, downward-directed angles or hooklike processes of the squamosals, mentioned above, extend some 23 mm. below the lower borders of the opisthotics. The boundaries of the supraoccipital are not distinct; it seems to be a small, approximately triangular element forming a considerable portion of the upper border of the foramen magnum. Its greatest width is probably not over 31 mm. The exoccipitals and opisthotics are thoroughly co-ossified. The *exoccipitals* form the lateral and probably most of

[1] *Ibid.*, p. 92.

the lower border of the foramen magnum. The part they take in the formation of the occipital condyle, I take it, is slight. The *opisthotics* are rather stout and widen considerably toward their lateral extremities. Just before they reach the hooklike processes of the squamosals their vertical measurement is about 38 mm. From this point they rapidly taper off in each direction. They form the lower and apparently the inner borders of the post-temporal vacuities. The relations of the *pterygoids* with the quadrates seem to be much the same as in *M. planirostris*, according to McGregor's interpretation. Their lateral extent on the posterior surface of the skull seems to be comparatively slight. There is no possibility of their taking part in the borders of the quadrate foramina as they do in *Belodon*[1] (*Mystriosuchus*) *plieningeri* H. von Meyer sp., according to F. von Huene.[2] In the specimen herein described the lower lateral portion of the quadrate was broken away, leaving, however, a good impression of the inner surface. From this and the natural mold in the matrix, of the outer surface, and with the aid of a separate quadrate found in the same locality, an accurate restoration was possible. It has an articular face of about 75 mm. lateral extent. There is an offset in the quadrate along a vertical line through the inner limit of this articular face. From this offset it extends toward the median line in a thin plate. Its inner edge meets the pterygoid and the two floor over a depression, the outline of which is a parallelogram with the greater diagonal directed upward and inward. This depression is bounded above by the opisthotic, on the outer side by the raised portion of the quadrate, below by the raised barlike portion of the cojoined quadrate and pterygoid, and on the inner side by the exoccipitals and the basisphenoid.

THE MANDIBLE

The mandible of this species is massive throughout. This is especially noticeable in the posterior symphysial region; here the

[1] For an explanation of the substitution of *Belodon* for *Mystriosuchus* see the appended list of genera and species.

[2] "Beiträge zur Kenntnis und Beurteilung der Parasuchier," *Geologische und palaeontologische Abhandlungen*, Neue Folge, Band X, Heft I (1911), p. 8, Fig. 3.

rami have a thickness of 40 mm. The smallest width of the mandible is 41 mm., at a point about 60 mm. back of the anterior end. From here the width increases gradually to about 90 mm. at the posterior end of the symphysis. The anterior end is considerably enlarged for the reception of the large terminal teeth. The greatest width of this expanded portion is 73 mm. The lateral margins are upturned so as to form a very conspicuous anterior-posteriorly directed groove along the median line. This is 31 mm. wide at the top and is broadly excavated to a depth of 9 mm. From the tip of the mandible to the posterior end of the symphysis the length is about 430 mm. The total length is about 9[illegible] mm. The external maxillary fenestra is about 200 mm. long by [illegible] mm. wide, with rounded anterior and posterior borders. Only the *alveoli* remain in the lower jaw upon which to base a description of the teeth. About 36 of these are present now on one side. There were probably 10 more present in the fragment which is now missing above the anterior end of the external maxillary fenestra, making about 46 alveoli in all. Three of these are placed in the expanded upturned anterior portion of the jaw. They are circular in section, as are all the alveoli, and greatly enlarged. The anterior one on each side is 12 mm. in diameter, the next two are of about equal size and measure 16.5 mm. in diameter. If the supposition that the tooth shown in Fig. 8 is one of these anterior teeth of this or a similar species is correct, we may assign a length of some 45 mm. or more to the teeth of the terminal expansion. The two anterior alveoli which are separated by the median channel mentioned above are about 24 mm. apart. Between the first and second and the second and third teeth only a thin film of bone intervenes. Between the last tooth of the terminal, enlarged series and the following one, the fourth, there is a space of about 15 mm. The alveoli back of the three large anterior ones vary considerably in diameter, ranging from 8 mm. to 16 mm. In a general way they increase in size posteriorly, but this is irregular, some of the largest being placed near the middle of the row. In consequence of this irregularity in the size of the alveoli there is an irregularity in the intervening spaces. These spaces range from a thin film to about 8 mm. in width, the greater spaces being, in general, in the anterior portion of

the mandible. This irregularity of size and spacing of the alveoli is probably largely due to the different ages of the teeth; that is, to loss and replacement of some. Well-developed, rounded ridges, such as are seen on the ventral surface of the rostrum, are present just within the lines of alveoli. In this specimen they are prominent anteriorly, combining near the anterior end and gradually spreading posteriorly. They become less prominent toward the posterior end of the symphysis and are entirely wanting before this point is reached.

The second specimen referred to in the preliminary notice of *Angistorhinus grandis*[1] would, it was hoped, when worked up, furnish a complete skeleton of that form. It was found, however, that the matrix was a hard, extremely tough sandstone and almost impossible to free from the bone, as the latter was in every case found to be softer than the matrix and extremely brittle. For these reasons a further attempt to prepare a skeleton has been abandoned, for the present at least. Aside from the skull, a portion of the posterior border of which is missing, and the lower mandible, which also lacks a little of the posterior extremities, only a single dorsal vertebra, several abdominal ribs, and quite an assortment of unassociated teeth are available for description. These serve well, however, to set the specimen off as a species distinct from *A. grandis*.

Angistorhinus gracilis Sp. Nov.

This skull, which I take to be that of an adult individual, probably had a length of about 985 mm., somewhat greater than that of the specimen of *A. grandis*. The greatest length preserved, which does not quite include the quadrate, is 920 mm. This is somewhat greater than the same measurement of *A. grandis*, but in spite of this the latter is the more massively built skull, a point that is especially noticeable in the form of the mandible to be discussed farther on.

OPENINGS OF THE SKULL

The supratemporal fenestrae are missing, but from the close resemblance of the skull to *A. grandis* in other respects these open-

[1] *Jour. Geol.*, XXI (1913), 190.

ings were probably about the same in both forms. That is, the parietal-squamosal arcades are well devleoped and in the same plane as the dorsal surface of the cranium. The upper posterior border of the lateral temporal fenestra is also wanting. Its width of 80 mm., however, which is greater than that of *A. grandis*, would indicate at least an equally large fenestra in *A. gracilis*. Contrary to what might be expected from this greater width of the lateral temporal fenestra, the bar of bone left between it and the orbit is much more massive than in *A. grandis*. The orbits are somewhat smaller than in this latter genus, about 45 mm. wide and 75 mm. long. Some 30 mm. below the orbit, in the jugal bar, midway between the lateral temporal fenestra and the antorbital vacuity, is a large, well-defined foramen, probably for the enervation of the powerful cheek muscles as in the modern *Gavial*. This was not noted, but probably was present in the specimen representing *A. grandis*. The nares are elevated on a considerable prominence, about 31 mm. above the general level of the roof of the cranium. As in *A. grandis*, they are separated by a thin median partition that does not rise to the same elevation as the posterior and lateral borders of the nares. Thus the nares form, in a way, a common opening about 35 mm. wide and 55 mm. long. The exact length cannot be told, as the anterior borders are weathered away. The state of preservation is such that few of the sutures are distinguishable. Hence a description of all the separate elements of the skull is impossible. Special attention was paid to the region bordering the nares in the hope that the relations of the septomaxillary bones could be determined. With but a few exceptions the students that have worked with the phytosaurs have overlooked these elements. According to F. von Huene they are present in all of the phytosauria. To quote:[1] "Vor dem Nasenlöcher schliessen die Nasalia nicht zusammen. Hier tritt ein ungewöhnlicher Knocken, das Septomaxillare (bei allen Phytosauriern) dazwischen."

A diligent search on the specimen representing *A. grandis* and also *Palaeorhinus bransoni* failed to show these elements. However, one is forced to conclude that the failure to locate the sutures of such elements is due to the state of preservation of the specimen

[1] *Op. cit.*, p. 11. The reference is to *Belodon* (*Mystriosuchus*) *plieningeri*.

and not to their absence, as the septomaxillary bones are apparent in the specimen of *A. gracilis.* In all probability the region about the anterior border of the nares in many of the phytosaurs was always more or less cartilaginous. In many of the specimens that the writer has studied, either from figures or from the original, this anterior border is ill defined, extending forward in a sort of slit. In *A. gracilis* the septomaxillae evidently unite along the median line and form the anterior border of the nares. Their extent in front of the nares along this line is 45 mm. What part they take in the median septum in separating the nares cannot be determined. Their greatest width is about 27 mm. on a line through the anterior border on the nares. In front the premaxillae dovetail into them and form their anterior border. The nasals meet the premaxillae far forward along the lateral border of the septomaxillae and form the lateral boundaries of those elements. The relations of these various elements are diagrammatically shown in Fig. 4.

The rostrum in this species is much more slender than that of *A. grandis.* From the tip to the anterior border of the nares is a distance of 670 mm., or about 610 mm. to the anterior border of the antorbital vacuity. In *A. grandis* these measurements are 650 mm. and 585 mm. respectively. The width of the rostrum just back of the terminal expansion is 34 mm., at about midlength 44 mm., and at a point 140 mm. in front of the antorbital vacuity, 56 mm. These respective measurements in *A. grandis* are 42 mm., 65 mm., and 94 mm. In the latter species the rostrum, seen in a lateral view, tapers down much less abruptly from the cranium proper than in the former. The terminal expansion of *A. gracilis* starts abruptly at a distance of 68 mm. from the end of the horizontal portion, increasing from a width of 34 mm. to 51 mm. This width it maintains, with only a very slight increase, to the tip. The extremity is bent abruptly downward and extends approximately 44 mm. below the ventral surface of the rostrum. The rostrum as a whole is curved, concave upward, but this is probably a result of deformation. As seems to be the usual case in phytosaurian remains, most of the teeth have dropped from the alveoli. Only the first, second, and fourth of the right premaxilla teeth are preserved in this specimen but these are entire and well preserved.

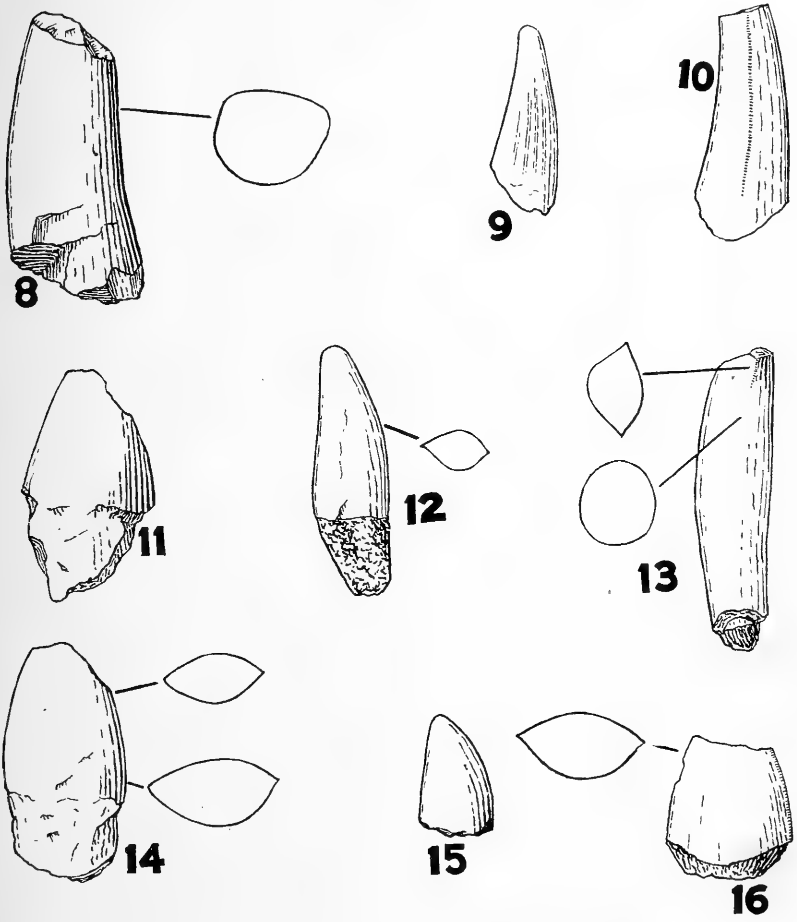
The first and second are in the down-curved tip of the rostrum. The fourth is probably the last of the long rakelike teeth, for it is the last tooth in the anterior expanded portion. The first tooth, the anterior one, is nearly complete, about 35 mm. long and 9 mm. in diameter at the base. This is probably a substitute for an earlier tooth, as the section of the corresponding tooth on the opposite side, which is broken off close up to the surface of the rostrum, indicates a tooth as large as those that follow. The second tooth is considerably larger and still preserves the sharp point. It is 78 mm. in length and has a diameter of 12 mm. at the base. The third tooth is broken off at the base but the section indicates a size for it equal to the fourth. This tooth is 60 mm. long and 9 mm. in diameter. The section of all of these anterior teeth is round with no indications of trenchant edges or fluting. The space between these teeth of the expanded anterior portion is somewhat greater than those between the remaining teeth. Between the two anterior teeth is a space of fully 18 mm. Between the first and second, the second and third, and the third and fourth the spaces are approximately 12 mm., 9 mm., and 10 mm., respectively. The number of teeth in the rostrum back of these cannot be told accurately. It is probably slightly greater than in *A. grandis*, however, as there are about 38 in the space from the tip to the anterior border of the antorbital vacuity in the former and 34 in the space of the latter. This is inclusive of those of the expanded anterior portion. In *A. gracilis* the alveoli back of the expanded portion are subequal, averaging about 10 mm. in diameter, and all are round. There is a slight increase in diameter from the anterior end of the series backward. The spacing is rather irregular, varying between 4 mm. and 9 mm. While the crowns of none of these posterior teeth are preserved in place, they can be described with some degree of certainty. From the locality where the skulls herein described were collected nearly all the vertebrate remains were phytosaurian. Many loose teeth were picked up as well as several associated with skulls. There is no group of animals from this locality other than the phytosaurs to which the teeth can be ascribed with certainty. A selected series of these teeth is shown in Figs. 8–16. One specimen (not here figured), a fragment of a skull in which some of the

teeth of the margin below the antorbital vacuity are still preserved, forms a part of the collection. This fragment is undoubtedly *Angistorhinus* sp. The crown of one of these teeth is well preserved and is almost identical with that shown in Fig. 16, which is one of several teeth found loose in the matrix close to the skull of *A. gracilis*. The crown of this tooth is 25 mm. long, 20 mm. wide at its greatest width, which is just above the alveolus, and at this same point has a thickness of about 12 mm. The anterior and posterior edges are sharp and very finely serrate, there being three serrations to one millimeter. Of the other teeth figured all but that in Fig. 9 have sharp anterior and posterior cutting edges with serrations numbering from fifteen to twenty in five millimeters. In that one, while the anterior and posterior edges are not sharp, the section is oval and it is further differentiated from the others by indications of fluting. In general it resembles the unerupted tooth seen in a section through the rostrum of *A. grandis* mentioned above. Just where this tooth found its place in the dentition cannot be told. It is the presumption, however, that it is one of the anterior teeth. While the ends of any series that can be formed of these teeth differ widely, they grade into each other in such a way that one would probably not be justified in supposing that such a series could not be found in the jaws of a single individual. Fig. 8, for instance, might well be the anterior tooth in the lower jaw of *A. grandis;* Fig. 10 is much the same sort of a tooth, but somewhat smaller. Fig. 12 is probably one of the premaxillary teeth, Figs. 11, 14, and 16, the middle or posterior maxillary teeth, and Fig. 15, one of the last in the series. The sections of Fig. 13 show well how the approximately circular root suddenly expands into the trenchant edged crown. Such variations as those above show the uncertainty encountered in basing species on a few isolated teeth with our present fragmentary knowledge of the dentition of these forms.

THE MANDIBLE

The posterior extremity of each ramus is missing, probably about 200 mm. The part present shows the same characteristics as does the skull when compared with *A. grandis*, viz., a greater length and more delicate build. The length of the symphysis is approximately 500 mm., considerably greater than in *Belodon*

(*Mystriosuchus*) *plieningeri* or *Lophoprospus* (*Belodon kapffi*). In these two latter forms, according to F. von Huene,[1] the length is 290 mm. and 280 mm. respectively. The end of the mandible swells rapidly at a point about 51 mm. from the tip into a disklike expansion with a width of some 68 mm. This terminal expansion



FIGS. 8–16.—Phytosaurian teeth, presumably from the genus *Angistorhinus*. Figs. 8 and 10, teeth of the anterior terminal expansion of the lower jaw? Fig. 9, probably one of the anterior premaxillary teeth. Figs. 11, 14, 15 and 16, posterior maxillary teeth? Fig. 12, probably one of the anterior premaxillary teeth. Fig. 13, root and part of crown of one of the posterior teeth, showing the sudden transition from round to compressed section. All a little under natural size.

[1] *Op. cit.*, p. 27.

differs from that of *A. grandis* in that the upper surface is a plane instead of being deeply grooved along the median line. Below the expansion there is a considerable convexity for the accommodation of the roots of the large terminal teeth. Just back of the expansion the mandible measures 29 mm. and has a thickness of about 17 mm.

Posteriorly the width increases gradually to 99 mm. at the posterior end of the symphysis or about 9 mm. more than the same measurement in *A. grandis*. It measures 25 mm. in thickness at this point, while this measurement is 41 mm. in *A. grandis*. These differences of measurement of the posterior end of the symphysis may be partially due to crushing in the former specimen but there is little evidence of this. Of the lower jaw none of the teeth is complete; the few that remain are broken off close up to the mandible and add little to our knowledge of the dentition except as to the relative size and spacing of the teeth. The alveoli are all more or less circular in outline but vary somewhat in size. In the expanded portion of the anterior end are placed three large teeth of approximately equal sections. They are apparently antero-posteriorly compressed, but this is probably an oblique section, a horizontal section through an outward-directed tooth. The average antero-posterior diameter is about 10 mm., considerably less than that of *A. grandis*. The two anterior teeth are separated by a space of 24 mm., while the other teeth of the expanded portion are crowded close together. The four teeth of the down-turned extremity of the rostrum evidently met close in front of the lower mandible, thus causing its large terminal teeth to close three on either side of the rostrum. The alveoli immediately back of the expansion are considerably smaller than the anterior ones and somewhat smaller than the teeth in the same region of *A. grandis*. They average about 5 mm. in diameter and are separated by a space of about 6 mm. From 4 mm. they increase quite regularly to 8 mm. or 9 mm. in the posterior ones and all are approximately 6 mm. or 7 mm. apart. In this species the teeth of the lower mandible are more numerous than in *A. grandis*, about 49 on each side.

THE VERTEBRAE

The single vertebra freed from the matrix is a little distorted and has lost the upper part of the spine but it shows the essential

features well. It is a dorsal vertebra, one of the most anterior. It corresponds in a general way with those of the same region in *Rhytiodin carolinensis* Emmons and may therefore be compared with one of the anterior thoracics of that form figured by McGregor.[1] The vertebra of *Angistorhinus gracilis* is considerably the larger and the proportions are somewhat different. The centrum is very much constricted laterally, measuring but 29 mm. in the middle. The posterior face of the centrum is approximately circular, measuring about 54 mm. in diameter. The transverse diameter of the anterior face is somewhat greater than the vertical diameter, due to the enlarged articulation for the tuberosity of the rib. These diameters measure 68 mm. and 69 mm. respectively. The diapophyses form heavy transverse processes extending horizontally at the level of the neural canal and expanding distally. They are approximately triangular in section with one vertex directed downward. The centrum is considerably excavated below the base of the diapophyses, leaving them supported for a short distance laterally by two thin buttresses which are confluent below with the anterior and posterior rims of the centrum. The capitular articulation (if it be the true articular surface) is slightly concave. The articulation for the tuberosity is low down on the anterior border of the centrum, is convex, and as large as the capitular articulation. The posterior zygopophyses are missing. The anterior zygopophyses have a considerable anterior prominence. They are large, are directed up and inward at an angle of about 45° from the horizontal, and, though this measurement may be exaggerated by distortion, measure 44 mm. between the center of the articular faces. The spine is broken off about 31 mm. above the neural canal. The total height of the vertebra must have been at least 150 mm., however, and probably more. The following table of measurements will serve to show some of the differences in the two forms:

	Rhytidodon carolinensis	*Angistorhinus gracilis*
Length of centrum	38 mm.	45 mm.
Height of centrum	42 mm.	54 mm.
Total height	120 mm.	(at least) 150 mm.
Width across diapophyses	104 mm.	156 mm.

[1] *Op. cit.*, p. 66, Figs. 15, 15*a*, and Pl. 8, Figs. 10, 10*a*.

VENTRAL RIBS

Weathered out about the specimen were many fragments of ventral ribs. Several of these ribs have been restored and all show a length of 260 mm. or more. Most of them are almost straight for half their length, whence they curve gently to the extremity. At the middle they are somewhat flattened, having greater and lesser diameters of about 16 mm. and 12 mm. respectively. They taper slightly to the extremities and here the section is almost round. These are probably lateral ribs of the ventral series. One of the ribs differs considerably from the others in the amount of curvature. It is broadly curved at the center with the extremities bent nearly at right angles. This is possibly one of the median ventral series.

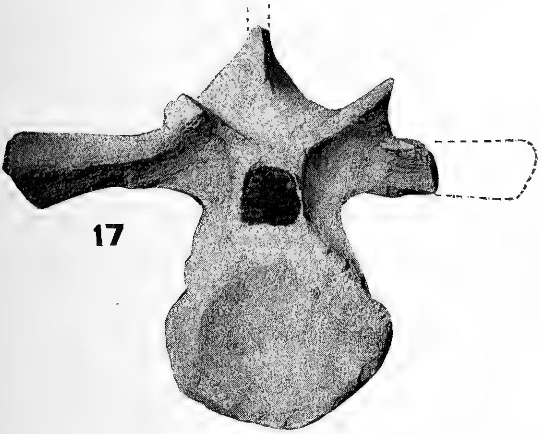
COMPARISON OF *A. grandis* AND *A. gracilis*

A brief review of some of the salient features of these two specimens will serve to show the distinctness of the two species:

1. Terminal expansion of rostrum taking place gradually and increasing in width to the tip.	1. Terminal expansion of rostrum taking place suddenly and retaining the same width to the tip.
2. Rostrum 94 mm. wide at a point 140 mm. in front of the antorbital vacuity.	2. Rostrum 56 mm. wide at a point 140 mm. in front of the antorbital vacuity.
3. Terminal expansion takes place in front of third tooth.	3. Terminal expansion takes place in front of fifth tooth.
4. Greatest length of skull is 977 mm.	4. Greatest length of skull is about 985 mm.
5. Length from anterior border of orbits to tip of rostrum 755 mm.	5. Length from anterior border of orbits to tip of rostrum 750 mm.
6. Interorbital width 68 mm.	6. Interorbital width 68 mm.
7. Terminal expansion of lower jaw upturned along lateral margins to form a median antero-posterior grove.	7. Terminal expansion of lower jaw in a plane.
8. Rounded ridges along inner side of alveoli prominent.	8. Rounded ridges along inner side of alveoli not prominent.
9. Number of teeth in lower jaw about 46 on each side.	9. Number of teeth in lower jaw at least 49 on each side.
10. Diameter of alveoli of lower jaw, exclusive of the large anterior ones, ranging between 8 mm. and 16 mm.	10. Diameter of alveoli of lower jaw, exclusive of the large anterior ones, ranging between 5 mm. and 8 mm.

A COMPARISON OF THE GENUS *Angistorhinus* WITH OTHER GENERA

Angistorhinus is one of the largest forms known. The skull is longer than any other except perhaps that of *Lophoprosopus* (*Belodon*) *kapffi*. The most striking difference seen between these two forms is in the rostrum. In *Lophoprosopus* there is a high, vertical swelling or crest extending almost the entire length, while in *Angistorhinus* the rostrum is low and slender. In this form, too, the anterior extremity is abruptly turned down, much more so than



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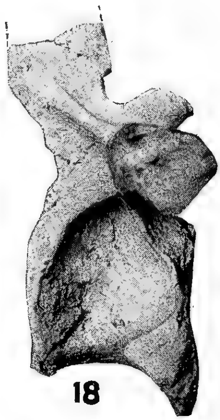

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FIGS. 17–18.—*Angistorhinus gracilis*, anterior and lateral views of one of the anterior thoracic vertebrae, about one-third natural size.

in *Lophoprosopus*. Another important difference is seen in the well-developed parieto-squamosal arcade lying in the plane of the dorsal surface of the cranium of *Angistorhinus* and the much-reduced, depressed post-temporal arcade of *Lophoprosopus* that gives the supratemporal fenestra the appearance of being open behind.

From *Mystriosuchus* the genus differs, not only in the development of the post-temporal arcade in the manner mentioned above, but also quite radically in the teeth. In *Mystriosuchus* there are 94 of these in the upper dentition and not more than 84 in the upper dentition of *Angistorhinus*. In the latter genus, as pointed out above, the posterior teeth are laterally compressed with sharp

anterior and posterior cutting edges. In the former genus, on the other hand, as shown by Dr. E. Fraas,[1] all the teeth are circular in section. In *Angistorhinus* the anterior end of the rostrum is much more abruptly down-turned and the downward extent is considerably greater than in *Mystriosuchus*. In the former genus, too, the posterior palatine foramina are much longer and more slit-like. In a lateral view the cranium proper of *Angistorhinus* is seen to make up a greater portion of the entire length of the skull than is the case in *Mystriosuchus*.

Rutiodon, of which *R. carolinensis* is the type, probably resembles *Angistorhinus* in general form more than any of the other genera. The differences found in the vertebrae have been pointed out above and it was shown that those of *Angistorhinus* were considerably the larger. The skulls, too, bear out this difference in size. The total length has not been determined for *Rutiodon* but the indications are that it was about that of *Mystriosuchus*, viz., 820 mm.,[2] while that of *Angistorhinus* is from 977 mm. to 985 mm. or more. The distance from the anterior border of the nares to the tip of the rostrum in the former is 510 mm.; in the latter genus the same measurement is 590 mm. Although a direct comparison of the dentition of the two forms is impossible, the teeth of *Rutiodon* at the middle of the rostrum seem to be considerably larger than those of *Angistorhinus*. The post-temporal arcade of the latter is greatly developed (in this it seems to differ from all other genera except perhaps *Palaeorhinus*) and the posterior palatine foramina are quite different. In the former genus they are almost round and probably not over 15 mm. in diameter, while in *Angistorhinus* they are about 60 mm. long and about 8 mm. in diameter.

Besides the much more anterior position of the nares in *Palaeorhinus* and the but slightly down-curved tip of the rostrum, this form differs from *Angistorhinus* in that the skull is much smaller and the slender rostrum makes up but about one-half the entire length. *Palaeorhinus* also has a less backward extension of the upper posterior border of the skull, much less developed "squamosal hooks," and the opisthotics are less massive, thinner, and more spatulaform than in *Angistorhinus*. The anterior extent of the

[1] *Op. cit.*, p. 16.
[2] McGregor, *op. cit.*, p. 58.

pre- and parasphenoid and of the inter-pterygoid vacuity is apparently considerably less in the former genus, and the pterygoid process of the basisphenoid is less massive and more separated from the greater wings than in the latter. The number of teeth in the upper mandible of *Palaeorhinus* is about 72, while in *Angistorhinus* they number 84. In all probability the post-temporal arcade is somewhat the same in both forms. Of this more will be said later in the special reference to *Palaeorhinus*.

The anterior part of the rostrum of *Mesorhinus fraasi* is missing. The rostrum therefore affords little material for comparison. What remains of it, however, suggests a form between *Angistorhinus* and *Lophoprosopus*. The external nares are much more anterior in *Mesorhinus* than in *Angistorhinus*, and the former, according to Jaekel, possess a parietal foramen. There is also apparently a difference in the elements about the anterior border of the nares in the two forms. Jaekel says:[1]

> Zwischen den vordern Enden der Nasenlöcher stehen zwei vertikale schmale Knochenleisten, die ich unbedenklich als Teile der Prämaxillaris angesprochen haben würde, wenn nicht Herr v. Huene ähnliche Gebilde bei anderen Belodonten eben als besondere Elemente, als Septomaxillaria beschrieben hätte.

F. von Huene, however, in a later paper, considers these as the ends of the premaxillae.[2] If we accept the interpretation of the latter, the difference between the two forms in this respect is evident. In *Angistorhinus* the septomaxillary bones seem to unite in the median line in front of the nares and exclude the premaxillae from that opening, while the premaxillae of *Mesorhinus* supposedly form the anterior and much of the lateral borders of the nares. While the parieto-squamosal arcade seems to be developed somewhat similar in the two forms, if one may judge from the figures of *Mesorhinus*, its posterior extent is not so great in the latter.

Palaeorhinus bransoni Williston

This genus and species was briefly described by Dr. S. W. Williston in 1904.[3] In 1907, after a detailed study of the type specimen, Mr. J. H. Lees published a much fuller description of

[1] *Op. cit.*, p. 201. [2] *Op. cit.*, p. 50. [3] *Jour. Geol.*, XII (1904), p. 696.

this form. Several new features were presented in this paper. To quote:

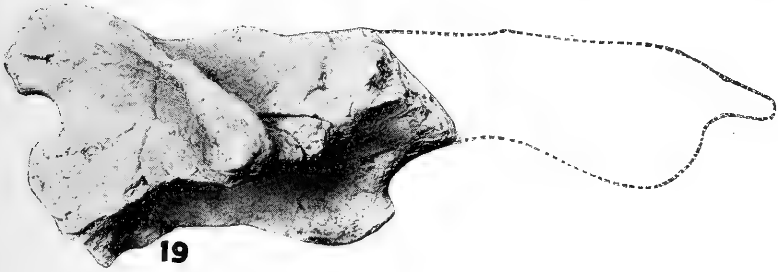
> The new features which the present study has disclosed may be here summarized: The presence of the otic capsule and its relations and dimensions have been delineated. The sutural relations of the palatines, pterygoid, and vomer have been more clearly delineated and the unsuspected posterior extension of the latter elements as far as the presphenoidal vacuity demonstrated.

And again (p. 147):

> In *Mystriosuchus* the parietals extend backward between the supratemporal openings at the level of the top of the skull for a little less than one inch, while in *Palaeorhinus* they extend in this direction two and one-fourth inches, are widely separated as already described, and meet the squamosals posteriorly to inclose the supratemporal at the upper level of the skull.

With these particular things in mind, viz., the posterior extent of the vomers, the otic opening, and the parieto-squamosal arcade, the writer has made a careful examination of the material representing this form. The skull is very much crushed, especially the palate, and few of the sutures are discernible. As the specimen is now prepared the articular face of the quadrate is turned almost directly outward instead of down, thus greatly exaggerating the posterior width of the skull. Basing the outline of a posterior view of *Palaeorhinus* on either *Mystriosuchus* or *Angistorhinus*, the greatest width would probably be not over 240 mm. in an uncrushed condition, or but little if any greater than that of *M. planirostrus*. In 1911, in a paper cited above, Jaekel mentioned the peculiar relations and development of the vomers in *Palaeorhinus* and presented figures showing what he considered a more likely interpretation of the palate. Jaekel's interpretation is certainly more phytosaurian. However, with the present state of the specimen one can only say that it is probably not justifiable to consider the vomers as extending back and forming the anterior border of the "presphenoidal vacuity." Concerning the otic capsule little can be said. The opening is present on the right side of the specimen; the left side is restored. This region was restored in *Angistorhinus grandis* from the impression of the inner surface of the bone. A thin film of the bone was still present and showed no indications of such an opening. Indeed, to the writer's knowledge, *Palaeorhinus*

is the only form in which the opening has been noted. In conversations with him, Dr. Williston has expressed the opinion that the opening is accidental in this case. One should probably not lay too much stress on its presence in this form till it has been found in other members of this genus. The parieto-squamosal arcade is crushed and largely broken away, giving an appearance to the supratemporal fenestra much like that of *Mystriosuchus planirostrus*. Whether this is the true condition or whether the form is like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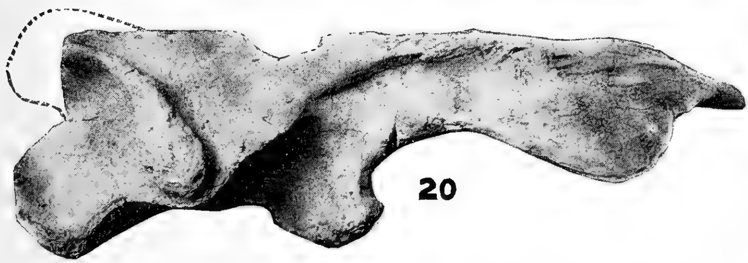
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FIGS. 19–20.—Ilia of a new, unnamed form. The ilium shown in Fig. 19 was referred by Lees to *Palaeorhinus bransoni*. Both figures about one-half natural size.

that of *Angistorhinus* as described above is not certain. The writer is inclined to agree with Lees, however, in assigning to it the form of the latter, but this view should probably also be held as tentative.

It has been impossible to ascertain the association of the ilium figured with the skull of *Palaeorhinus*.[1] Material has recently come into the hands of the writer which shows that the ilium is not that of *Palaeorhinus* as Mr. Lees supposed, but belongs to an entirely different group of reptiles. Fig. 19 is of the same ilium figured with

[1] Lees, *op. cit.*, Fig. 8.

Palaeorhinus, with slightly different orientation to bring out its similarity with another specimen, Fig. 20. This second specimen is somewhat crushed laterally and is slightly smaller and more slender than the first. They probably represent two different species but the figures indicate that they are generically the same, or at least should be so considered till evidence to the contrary is forthcoming. A brief notice of this new form follows:

NEW GENUS?

The material upon which this form is based consists of several extremities (both proximal and distal) of limb bones, several vertebrae from various regions of the body, and many other fragments, besides the ilium figured above (Fig. 20). The specimen is a part of the collection of the University of Chicago expedition of 1904 and comes from the Popo Agie River region of the Trias of Wyoming.

One of the vertebrae has been freed from the matrix and shows characteristics not at all like those of the phytosaurs. It is a middle or anterior thoracic vertebra with the capitular articulation in a transitional stage in its elevation from the centrum to the arch. The centrum is spool- or hourglass-shaped with anterior and posterior faces slightly concave and is much constricted laterally along its midlength. It is 54 mm. long and 10 mm. wide at the center (this width is probably lessened by pressure). The articular faces are slightly oblong in outline, about 32 mm. wide and 39 mm. high. The diapophyses are in the form of thin horizontal plates confluent anteriorly with the articular faces of the zygapophyses and widening regularly posteriorly. Suddenly, near the posterior ends, they thicken below into the articulation for the tuberosity. At this point the diapophyses are about 56 mm. from tip to tip. They are supported at the posterior end by two thin diverging buttresses, one directed toward the upper posterior articular face of the centrum but not reaching this point, the other confluent with the capitular articulation on the upper anterior articular face of the centrum. The distance between the articular faces for the tuberosity and that of the capitulum is some 30 mm. The total height of the vertebrae is 96 mm. The upper portion of the spine is

missing, however—probably no small portion. At the base the spine is thin, not over 5 mm. in thickness, and has a width of 43 mm.

Several forms have been described from the region where these remains were found, all based on more or less fragmentary material.[1] The indications all point to their distinctness from the form herein described, but as yet not enough is known to warrant its being given a new generic name. The writer expects soon to make a series of thorough collections in the western Trias and hopes to add to the knowledge of this and other forms that are little known. A fuller description of the material here mentioned and a discussion of its relationships will follow.

All who have worked with the phytosaurs have felt, I am sure, that we have been too zealous in our efforts to place all the fragments that have been referred to this group in the various schemes of classification. Too often we have proposed new species and even new genera for a few tooth fragments or a few imperfect limb bones.

The following list consists of those forms that, in the writer's opinion, have been fairly well established. In the arrangement of these forms no attempt is made to show the genetic relationship.

1. *Phytosaurus Jaeger*[2]

P. cylindricon Jaeger

In 1861 H. von Meyer attempted to show that *P. cylindricon* and *P. cubricon* Jaeger were in reality described from different regions of the jaw of one individual and that this form was the same as his *Belodon kapffi*.[3] This being the case, the name *Phytosaurus* has priority, and McGregor has so used it (1906) referring to that genus the forms that had previously been placed in the genus *Belodon*. The material upon which the genus *Phytosaurus* was based, however, is in such an exceedingly poor state of preservation

[1] For a description of *Dolichobrachium gracile* Williston, *Eubrachiosaurus browni* Williston, and *Brachybrachium brevipes* Williston, see *Jour. Geol.*, XII (1904), 688–94.

[2] G. F. Jaeger, *Über die fossil Reptilien welche in Württemberg aufgefunden worden sind.* Stuttgart, 1828.

[3] "Reptilien aus dem Stubensandstein und Keuper," *Palaeontographica*, VII 253–346.

that one can never safely refer other material to it.[1] The species *P. cylindricon*, then (the form first described), is the type and only species of this genus, and the form described by von Meyer as *Belodon kapffi* should be considered distinct.

2. *Belodon* von Meyer

    B. plieningeri von Meyer

    B. (*Mystriosuchus*) *rutimeyeri* F. von Huene

The genus *Belodon*, of which *B. plieningeri* is the type, was proposed in 1842[2] for a form with a slender rostrum and laterally compressed, arrowhead-shaped posterior teeth.

3. *Mystriosuchus* E. Fraas

    M. (*Belodon*) *planirostris* H. von Meyer

This form differs from the genus *Belodon* chiefly in that the teeth of the former are all round in cross-section.

4. *Rutiodon* Emmons[3]

    R. carolinensis Emmons

    R. manhattanensis von Huene

The writer has followed McGregor in uniting in the first species most of the many specimens that have been described from the Trias of the eastern United States under the following names:

Clepsysaurus pennsylvanicus I. Lea.
Centemodon sulcatus I. Lea.
Omosaurus perplexus J. Leidy.
Palaeosaurus sulcatus E. Emmons.
Palaeosaurus carolinensis E. Emmons.
Compsosaurus priscus J. Leidy.
Clepsisaurus leaii E. Emmons.
Eurydorus serridus J. Leidy.
Belodon carolinensis E. D. Cope.
Belodon priscus Cope.
Belodon lepturus Cope.
Belodon leaii Cope.
Rhytiodon rostratus O. C. Marsh.
(?) *Beldon validus* Marsh (doubtful, never figured).

[1] Although the writer has not seen this material, he has been informed by no less an authority than Dr. S. W. Williston, who has recently had the opportunity to examine the type remains, that the preservation is such that one can in nowise be certain of its generic identity.

[2] Von Meyer, *Neues Jahrb. f. Mineralogie.*

[3] Geological report on the midland counties of North Carolina.

5. *Lophoprosopus* gen. nov.
   L. (*Belodon*) *kapffi* von Meyer
   (?) *L. buceros* Cope[1]

The generic value of the form represented by *Belodon kapffi* von Meyer is well established. The name *Lophoprosopus* is here substituted because, as pointed out above, both *Belodon* and *Phytosaurus* are preoccupied.

In the past several species have been based on very doubtful material and referred to the group represented by *L. kapffi*. These deserve, perhaps, brief mention here:

Belodon (*Zanclodon*) *arenaceus* E. Fraas, a fragment of a mandible that resembles *L. kapffi* and may well belong to that species.

Belodon ingens E. Fraas, a single skull very similar to that of *L. kapffi* and probably the same.

Belodon scopax E. D. Cope. Judging from the snout this form may well belong to the genus *Palaeorhinus*.

Belodon superciliosus E. D. Cope. This species is based on a few skull, teeth, and scute fragments and is very doubtful.

Heterodontosuchus ganei F. A. Lucas. The fragment of a mandible upon which this species was based was originally described as a Triassic crocodile. According to McGregor[2] we may expect to find this to be identical with *L.* (*Phytosaurus*) *buceros* when better material is at hand.

6. *Palaeorhinus* Williston
   Palaeorhinus bransoni Williston
   (?) *P. scopax* (see the reference to this species under the genus *Lophorhinus*)
7. *Angistorhinus* Mehl
   A. grandis Mehl
   A. gracilis Mehl

[1] Some question arises as to whether this form belongs here. To quote Dr. O. Jaekel (*Sitzungsberichten der Gesellschaft naturforschender Freunde*, No. 5, Jahrgang 1910, pp. 219–20): "Durch die Gesamtform des Schädels, besonders die Vorwölbung des pränasalen Schnauzenteils, die weit rückwärtige Lage der Nasen und die weite Vorstreckung der Squamosa-Ecke ist diese Form im Rahmen der *Phytosauridae* so deutlich gekennzeichnet, dass ich für sie daraufhin einen neuen Gattungstypus vorschlagen, und ihm den Namen *Metarhinus* geben möchte."

It is probably well to refer this form tentatively to the genus *Lophoprosopus* till further discoveries settle the point.

[2] *Op. cit.*, p. 94.

8. *Mesorhinus Jaekel*

Mesorhinus fraasi Jaekel

9. *Parasuchus* Lydekker

P. hislopi Lydekker

Doubtful genera and species:

Rylea platyodon E. von Huene gen., Riley and Stutchbury sp.

These few imperfect remains, consisting of incomplete limb bones, teeth, caudal vertebrae, etc., are in all probability phytosaurian, but the material seems to be too meager to receive a generic or even a specific name.

Episcoposaurus E. D. Cope

E. horridus E. D. Cope

E. haplocerus E. D. Cope

Little is known of either of these two forms and their reference to the phytosaurs is not entirely warranted.

The genus *Palaeochtinus*[1] Cope, under which were described *P. appalachianus*, *P. orthodon*,[2] and *P. dumblainus*,[3] was based on a few teeth that Cope took to be those of a large dinosaur. These resemble closely some of the teeth figured with *Angistorhinus* and other phytosaurs and may well be phytosaurian remains.

Incertae sedis:

Steganolepis robertsoni T. H. Huxley

While the fragmentary remains that represent this form may indicate their phytosaurian affinity, the relationship is not at all clear. In view of our meager knowledge of this form and also *Mesorhinus*, it hardly seems advisable at this time to remove the latter form from the *Phytosauridae* and place it with *Steganolepis* in the family Steganolepidae as F. von Huene has suggested.[4] Probably undue stress has been placed on the presence of a parietal foramen in *Mesorhinus*. Only a single incomplete and poorly preserved skull has as yet been found. This fact being considered, it would seem that the presence of the parietal foramen is in nowise definitely established.

[1] *Proceedings Amer. Philos. Soc.*, 1877, p. 182.

[2] "A Preliminary Report of the Vertebrate Paleontology of the Llano Estacado," *Geol. Survey of Texas, Fourth Ann. Rept.*, 1892, p. 15, Pl. 2, Fig. 1.

[3] *Ibid.*, p. 16, Pl. 2, Figs. 4–6.

[4] *Op. cit.*, p. 50.

The writer wishes to take this opportunity to express his appreciation for the privilege of studying the material in the University of Chicago collections and also to thank Dr. S. W. Williston for his generous aid.

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THE ORIGIN OF THE INCLUSIONS IN DIKES

SIDNEY POWERS
Geological Museum, Cambridge, Massachusetts

PART II

(*C*) *Many inclusions have risen.*—

Shelburne Point, Vermont: On Shelburne Point and on Nash's Point, near-by, a few miles south of Burlington, Vermont, are several inclusion-bearing dikes of great interest. They have been described by Hitchcock[1] and by Kemp and Marsters.[2] On both sides of Shelburne Point are outcrops of a 20-foot bostonite dike filled with angular inclusions of Middle Ordovician shale and red Cambrian quartzite. These inclusions have sharp edges and do not appear to have been noticeably altered by the bostonite. On Nash's Point is another bostonite dike, about 12 feet wide and vertical, with chilled margins 2 feet and 1 foot wide on the respective sides of the dike, and a mass of fragments in the middle. A part of the fragments are quite angular and a part decidedly rounded. They vary in size from a fraction of an inch to 4 or 5 inches in length, but Kemp and Marsters found one piece of norite 18 feet in diameter. They consist of garnetiferous hornblende schist, pre-Cambrian norite, quartz, grey sandstone, red Cambrian sandstone, Trenton shale, and black limestone, cemented by a bostonite groundmass. Kemp and Marsters add: "Under the microscope, sections of norite show plagioclase and garnet, all exhibiting the results of dynamic action. Sections of the red quartzite have the usual fragmental character, with the evidences of strain less developed." The presence of the norite and schist inclusions show that the dikes have come through an indefinite amount of the pre-Cambrian as well as through Cambrian sandstone and Ordovician limestone into

[1] "On Certain Conglomerated and Brecciated Trachytic Dikes in Vermont," *Proc. Amer. Ass. Adv. Sci.*, XIV (1860), 156.

[2] "Trap Dikes of the Lake Champlain Region," *U.S. Geol. Surv., Bull. 107.*

their present position in Middle Ordovician shales. On the end of Nash's Point a 10-foot sill of bostonite is filled with unaltered shale inclusions from the immediate walls.

Kemp and Marsters have the following suggestions to offer concerning the manner in which the dikes acquired the inclusions:

> Two explanations may be offered for this rock. One, that it has been intruded in a line of previous faulting and attrition, which has broken up the walls and left loose material to be gathered up by the intrusive magma. This explanation has the greater weight with the writers. The other is that it represents only the upper portion of a dike and thus contains the float material which the advance of an intrusive body, that forced its own passage, would naturally gather from its walls. The lack of such inclusions in the neighboring dikes may be due to the fact that their tops have been eroded.

The difficulty with the first explanation is that there is no evidence of any faulting with brecciation at the particular places where the dikes came up. Moreover, how did the sill which follows a sharp fold in the shale obtain its inclusions by this hypothesis? It would appear extraordinary for faulting to occur in both a vertical and nearly horizontal position (the shales having a low dip) and to have brecciation occur only in these particular cases. The second explanation is partly in accord with the theory about to be proposed, but there is no reason why these should be the top of the dikes, or why the dikes for some distance vertically should not be equally filled with inclusions.

The origin of the inclusions appears to the writer to have been the shattering of the walls of the dike by the intrusive magma as it ascended through the pre-Cambrian, Cambrian, and Ordovician. Little corrosion could take place in fragments caught in such a magma and shot upward to their present position where they would be quickly chilled because of the inadequate supply of heat and lack of time. The invaded rocks would be comparatively cold when ripped off, and would therefore have a chilling effect upon the surrounding dike-magma in a 20-foot dike. Kemp and Marsters found the bostonite to be porphyritic, with a fine groundmass showing flow-structure around the inclusions. The rounded character of some of the inclusions is probably due to mechanical causes, principally the friction of the blocks against the walls of the dike

and against each other in ascent. The thickness of the sedimentary series in the region is about 12,000 feet from the top of the pre-Cambrian to the horizon of the outcrop of the dike. Hence, the pre-Cambrian norite inclusions must have risen over two miles and their edges would be necessarily rounded.

The abundance of the inclusions in two dikes and a sill, while other near-by dikes of the same composition and age do not show any inclusions at the present exposure, is not sufficient evidence to prove that all of the dikes do not contain inclusions in place. The shattering of the walls of some of the dikes may have been a purely accidental phenomenon, due to the way in which the fissures formed through which the magma came. If the passage were through a vertical fissure with straight unbroken walls, no inclusions would be expected.

Mancos, Colorado: Professor Kirtley F. Mather, of Fayetteville, Arkansas, has kindly contributed the following account of the inclusions in dikes near Mancos, Colorado:

In the San Juan region of southwestern Colorado, about 8 miles southwest from the town of Mancos, there is a small area of igneous rock which is locally known as "the Blow-out." The Mancos River at this point is flowing in a steep-sided youthful valley cut through the Mesaverde formation, which caps the neighboring plateaus, and far into the Mancos shale. "The Blow-out" is situated high up on the eastern slope of the valley in the midst of the maturely dissected shales. It consists of a conical mound about a quarter of a mile in diameter and two or three hundred feet in height, composed of mediosilicic extrusive volcanic material, flows, tuffs, and breccias. Cutting the volcanic cone and the surrounding country rock there are several subsilicic dikes which vary in width from an inch to 4 or 5 feet. The more prominent of these dikes can be traced for many hundred yards outward beyond the boundary of the central cone. Both the extrusive and intrusive rocks contain an abundance of inclusions of several different types of rock materials.

The extrusive volcanic rock varies considerably in its composition in different parts of "the Blow-out," but typically it is an andesite consisting of plagioclase feldspar, about $Ab_{55}An_{45}$, accompanied by biotite, magnetite, and augite, imbedded in a glassy matrix. The dike rock likewise is quite variable in composition, but it consists essentially of phenocrysts of biotite, augite, olivine, and plagioclase, imbedded in a groundmass of glass, pyroxene, felty plagioclase, magnetite, and biotite. The percentage relations of the minerals in the phenocrysts is notably different in different parts of the same dike, but usually the rock is rich in biotite and poor in plagioclase. The latter is little, if any, more calcic than that in the extrusive rocks.

Inclusions in the dike-rocks are extremely numerous and vary in size from tiny fragments up to masses nearly 3 feet in length. They consist of rounded, subangular, or occasionally angular, fragments of shale, sandstone, granite, metamorphosed limestone, and quartzite conglomerate. The inclusions of sedimentary rock are relatively abundant in approximately the same proportions as those in which the shales, sandstones, limestones, and conglomerates occur in the underlying sedimentaries—the Mancos shale, the Dakota sandstone with its basal conglomerate, the McElmo and LaPlata formations, etc. In the thicker dikes, shale inclusions are more numerous near the contact walls than toward the middle of the dike. The inclusions of sedimentary rocks are evidently fragments from the walls of the conduits through which the lava passed upward and from which they were torn by some process analogous to stoping. They all show the effects of their immersion in the molten mass with which they were surrounded.

The origin of the granitic inclusions is less apparent. No granites are known to outcrop within the drainage basin of the Mancos River, which includes much of the LaPlata Mountains northeast from Mancos. It is believed, however, that their origin must be similar to that of the other inclusions in the same dikes, and hence it is inferred that beneath the sedimentary rocks which outcrop in this region there must be a body of granite which was likewise cut by the dikes and from which fragments were torn by the ascending lavas in the same manner as those from the overlying strata. Such a granite mass might be either an intrusion into the sedimentaries, preceding the vulcanism, the effects of which are now displayed at the surface, or it might be a hill of pre-Cambrian or Paleozoic rock around and above which the Mesozoic sediments were laid down. These inclusions of igneous rock display a remarkable selective assimilation by the dike-magmas of certain of the minerals of the granite. This will be discussed in a paper now in preparation.

Aschaffenburg, Germany: In the Spessart region near Aschaffenburg there are a number of lamprophyre dikes, some of which contain inclusions of granite or of minerals derived from an augen gneiss. In the Schweinheimer kersantite dike, Thürach[1] has found inclusions 40 cm. in diameter, of a granite which contains orthoclase augen 5 cm. in length. Goller[2] has described quartz and orthoclase inclusions from similar dikes. The quartz fragments are rounded, and have a maximum length of 10 mm. The orthoclase fragments are less abundant and have a maximum length of 6 cm. These inclusions are scattered through the dikes. They

[1] "Über die Gliederung des Urgebirges im Spessart," *Geogn. Jahreshefte*, V (1893), 101.

[2] "Die Lamprophyrgänge des südlichen Vorspessart," *Neues Jahrb. f. Min.*, Beilageband VI (1889), 521 ff.

have come from the same underlying gneiss as the granite fragment. The selective action of the dike-magma in including fragments of only quartz and feldspar from the underlying gneiss is comparable to the case at Mancos. This fragmentation is favorable to the contention of Day, Sosman, and Hostetter[1] that the shattering of a siliceous rock is due principally to the expansive force of quartz at the inversion point of 575°C.

Somerville, Massachusetts: In Somerville and Medford, Massachusetts, are a number of exposures of diabase dikes, among which the Medford dike and the dikes of the Mystic River quarry carry inclusions. These dikes have variable mineralogical compositions. The Medford dike cuts Cambridge slate (probably of Permian age) and the older igneous complex of the Middlesex Falls. It varies in width up to 270 feet, and is probably over 4 miles in length. The Mystic River dikes cut the Cambridge slates, and are less than 10 feet in width. The inclusions in these dikes are chiefly composed of aggregates of single minerals, with vein quartz predominating. They are usually less than 6 inches in diameter and have well-rounded surfaces. Some of the quartz inclusions are fully 2 feet in diameter. A list of the inclusions found in the Medford dike includes[2]: quartz-diorite pegmatite, granite, graphic granite, diorite, quartzite, quartz schist, slate, quartz, feldspar, hornblende, and apatite. The inclusions in the Mystic River quarries are: altered biotite granite surrounded by a secondary rim of finer granite which has recrystallized with more femic constituents, diorite, pegmatite, quartzite, quartz, andesine feldspar (Ab_4An_3), hornblende, biotite, magnetite, apatite, zircon, and graphite.

The quartz inclusions are of especial interest and have been described by T. A. Jaggar, Jr.[3] These fragments have an irregular, clastic form. Sometimes the edges are rounded and sometimes very angular, with frequent embayments from magmatic corrosion.

[1] "The Determination of Mineral and Rock Densities at High Temperatures," *Amer. Jour. Sci.* (4), XXXVII (1914), 1–39.

[2] In making out these lists the writer has looked through the collections made by Professors Jaggar, Woodworth, and Palache, and is indebted to them for permission to publish the results.

[3] "An Occurrence of Acid Pegmatite in Diabase," *Amer. Geol.*, XXI (1898), 203–13.

A narrow-necked embayment, 3 inches in depth in the end of an inclusion 1 foot in length and 3 inches in width, has been noted by Jaggar. The quartz inclusions all appear to be surrounded by a wreath of augite prisms about 1 mm. wide, characteristic of quartz inclusions in basalt. This endomorphic reaction rim was found by Jaggar to consist of four zones between the diabase and the quartz: (1) diabase feldspar, (2) augite crystals, (3) potash feldspars, (4) micropegmatite.

The inclusions in these dikes appear to have been principally derived from rocks underlying the Cambridge slate. Very coarse diorite pegmatites, specimens of which have been found at Kidder Avenue in the Medford dike, and quartz veins seem to have furnished most of the material for the inclusions, the fragments of rocks being quite scarce, as would be expected in normal dikes shattering a few blocks off their walls during their ascent. The diorite pegmatite is not exposed in the vicinity of the dikes unless it is the same as that described by Jaggar near Arlington Heights.

These inclusions may have floated up in the dike-magma or they may have been forced up by it. Day *et al.* (*op. cit.*) have shown that the density of diabase glass is 2.763 and therefore heavier than the mineral xenocrysts. Whether the latter were formed by the differential expansion of the various minerals, a theory whose importance is emphasized by Goldschmidt,[1] or whether they were formed by purely mechanical action, is not clear. Numerous small xenocrysts of quartz and feldspar, from the vein quartz and diorite pegmatite, are found throughout the diabase, at least part of which have been broken off the corners of the inclusions.

Quartzose inclusions appear to be corroded frequently in igneous masses. A number of cases of reaction rims around quartz inclusions in extrusive rocks will be found described by Lacroix.[2] In the Globe district, Arizona, intrusive diabase contains numerous inclusions of vein quartz which are conspicuously corroded and embayed and surrounded by reaction rims of amphibole.[3] The

[1] *Die Kontaktmetamorphose im Kristianiagebiet* (1911), pp. 107–9.

[2] "Les enclaves des roches volcanique," *Annales de l'Académie de Mâcon*, X (1893).

[3] *U.S. Geol. Surv., Geol. Atlas*, Globe Folio (No. 111), 1904. Other diabase intrusions are described which contain many inclusions which have sunk.

Firth of Forth, Scotland, quartz diabases owe their free silica to solution of acid country rocks in normal diabase and contain corroded quartzose inclusions.[1]

Ogunquit, Maine: About 2 miles south of Ogunquit, Maine, several inclusion-bearing sills are exposed on the seashore halfway between Perkins Cove and Bald Head. The rocks of the region consist of thick-bedded slates, tilted into a vertical position with a uniform northeast strike, cut several miles inland by biotite granite. The slate is intruded by vertical trap sills of two or three generations, the older ones being porphyritic. The sills are mostly 5 feet or less in width, but one is over 50 feet wide. Most of the inclusion-bearing sills are of this older porphyritic type, which is seen under the microscope to be an augite-biotite kersantite. The sills occasionally cut across the bedding of the slates. They are younger than the frequent quartz veins which cut the slate.

In many of the sills in the vicinity there are a few inclusions. The latter are often 6 to 8 inches in length, with their longer axes parallel to the sides of the sill. In one sill, 2 feet, 10 inches wide, are two rounded inclusions of coarse graphic granite about 16 and 12 inches in diameter respectively, an angular block of granite 16 inches long and about 3 inches wide, with smaller subangular fragments of quartz and granite. There is a fused contact around all these inclusions. A 7-foot sill contains in one place many small subangular and rounded fragments of quartz, granite, syenite, and slate.

The principal inclusion-bearing sill is exposed for about 400 feet, showing a variation in width from 3 to 5 feet. It runs parallel to the stratification and has resisted erosion by the sea, while the slates on one side have been eroded away. The sill is cut by two younger dikes.

The character of the inclusions varies greatly. They are composed, in order of relative abundance, of a moderately coarse-grained light-colored syenite with green hornblende as the dark constituent; a coarse-grained granite resembling the syenite but containing free quartz; fine-grained pink and white aplitic granites which are probably biotite granites; fine-grained syenite seen in

[1] E. Stecher, *Tschermak's Mitt.*, IX (1888), 193.

thin section to consist principally of microperthite and oligoclase feldspars, with chlorite and other alteration products and a few grains of quartz; slate similar to the slate adjoining the sill; and vein quartz evidently derived from the veins which cut the slate. The size of the inclusions ranges from a few inches to 4 feet in length, in various parts of the sill. Thus, the inclusions in the part shown in the left-hand section of Fig. 1, where the sides of the sill are chilled, have an average size of 8 inches long by 4 inches wide. Nearer the sea, the margins of the sill do not show chilling, as

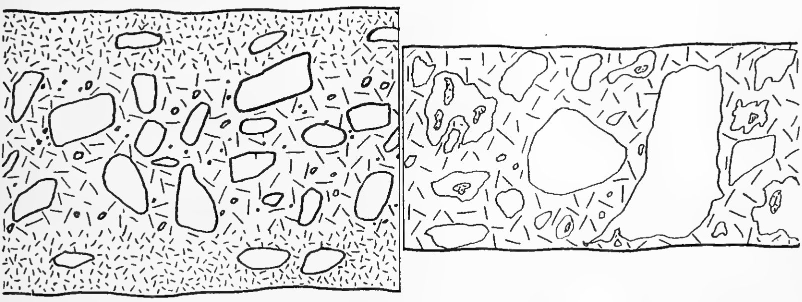

FIG. 1.—Two portions of the inclusion-bearing sill at Ogunquit, Maine. In the left-hand section the walls of the sill are chilled and the fragments show fused contacts but not resorption. In the right-hand section—about 100 feet away—the inclusions are being resorbed and their edges are indistinct. The width of the sill in the right-hand section is $3\frac{1}{4}$ feet.

shown in the right-hand section of Fig. 1, and here the average length of the fragments is about 1 foot. In the first area the inclusions have subangular to rounded outlines and tightly sealed contacts. In thin section the quartz from the aplitic granite can be seen to have migrated a sixteenth of an inch into the labradorite-augite-biotite matrix of the sill, but the feldspars from the coarse syenite have not migrated. In the second area, the inclusions all have corroded margins and the keratophyre may be seen to have worked its way into the coarsely crystallized syenite and granite. Blocks of the pink aplitic granite 3 and 4 feet long respectively and about $1\frac{1}{2}$ feet wide, with their longer axes orientated in some cases perpendicular to, and in other cases parallel to, the walls of the sill, where the latter is $3\frac{1}{4}$ feet wide, show that they have been

subjected to high temperature for a long period of time, as long tongues of granite run out into the dike-rock. In other cases merely remnants of inclusions over a foot in diameter are left, the greater part of the volume of the original block being filled with keratophyre mixed with the feldspars from the inclusion.

The origin of the inclusions of the igneous rocks must have been at some distance below the present exposure. No brecciation of the slate is observable at the surface, so it is thought that the inclusions of slate, which are often 2-foot cubes, and quartz, which are always small, also came from some depth, but not from such a distance as the igneous rocks. It is not known to the writer whether syenites, aplitic granites, and graphic granite outcrop farther inland or not. At Wells, 7 miles to the north, biotite granite is quarried.

The manner in which the inclusions of igneous rock, and possibly those of slate and quartz, were obtained is apparently by shattering, probably along the walls of the feeders of the sills. The resorption of the edges of the blocks must have been begun in depth where there was sufficient heat to assimilate a large part of an originally cubical block over 1 cubic foot in volume. However, as pointed out in other cases, if these blocks were shot up to their present position, the resorbed edges would be worn off; if they floated up, the resorbed edges would be partly preserved. Therefore, it is necessary to assume that the temperature of both the inclusions and the sill was nearly the same and that the corrosive action continued for some time after the blocks were at about their present levels, due to an additional supply of heat from below. This hypothesis is supported by the fact that where the margins of the sill are chilled, the inclusions in the marginal zones and also the more abundant inclusions in the center of the dike show fused contacts, but not extensive resorption. On the contrary, where there are no chilled margins, indicating that this part of the dike remained hot longer than the rest and then consolidated at one time, the inclusions all show resorption and assimilation and only skeletons are left of some of them. As the sill is of aschistic composition, such heating action is thought to be possible.

Rossland, British Columbia: During the C2 excursion of the International Geological Congress last summer, an inclusion-

bearing dike was observed by the writer at Rossland, British Columbia. The dike, which is known locally as the "White Dike," is 6 feet in width and consists of a mass of large bowlders imbedded in an augite kersantite matrix. Two types of bowlders were observed in the outcrop; one a white, fine-grained rock which was mistaken in the field for a quartzite, the other a bluish-grey rather coarse-grained rock with many small dark-brown patches. The texture of both is gneissic. A microscopic examination has shown that both types are anorthosite, the former consisting almost entirely of labradorite (Ab_{40},An_{60}), the latter of similar labradorite with small amounts of pyroxene, probably an orthorhombic variety with positive optical character. The pyroxene has in part altered to biotite and chlorite. Magnetite occurs in small amounts in both specimens. They both show the effects of some dynamic movement.

FIG. 2.—An exposure of the "White Dike" at Rossland, British Columbia, showing the abundance of inclusions. The sketch has been traced from a photograph. The scale is given by the width of the dike, which is 6 feet.

The inclusions vary in size, as shown in Fig. 2, the larger being 1 to 2 feet long. In shape they are partly angular, but in one of the surface exposures they are subangular or well rounded and do not have sharp edges, nor do the edges show any effect of corrosion. The inclusions weather out of the matrix, but slightly fused contacts are shown in the weathered material. The dike-rock consists of basic andesine or labradorite and augite phenocrysts in a matrix of labradorite, augite, biotite, and some magnetite.

The dike is in the extreme western portion of the Rossland district, about 5 miles north of the international boundary. It is shown on the Candian Geological Survey map of Rossland, at the western edge of the map, outcropping in a cut of the Great Northern Railway and at a flume 1,000 feet to the northwest. It cuts porphyritic monzonite in the first outcrop and augite porphyrite in the second. It has also been found in the 900-foot level of the Josie

mine, and it there presents the same appearance as on the surface and in the upper workings.[1] A similar dike has been found in the Columbia-Kootenay mine, but the inclusions are not as large as in the "White Dike."

The sedimentary rocks of the region consist, according to the map of the Boundary Creek mining district, of Paleozoic sediments, some of which are metamorphic, and Tertiary conglomerates and tuffs. The igneous rocks are of a number of types and represent several periods of intrusion. Gneisses of questionable age occur in the region, but no anorthosite has been reported.

Three explanations may be offered for the immediate origin of the inclusions:

1. The dike-magma may have come through a thick series of conglomerates of Paleozoic age and carried the pebbles and bowlders upward, dissolving their cement.

2. The dike-magma may have intruded Tertiary conglomerates, which have since been removed by erosion, the bowlders sinking in the molten dike-rock.

3. The dike-magma, when at some distance beneath the surface may have shattered off blocks of the rocks through which it passed, and carried them upward together with the blocks shattered in the formation of the fissure. The friction of the numerous inclusions against each other and against the walls of the dike would remove the angles and give the blocks rounded outlines.

These possibilities will be discussed in order.

1. If there were a thick series of conglomerates in the Paleozoic sediments, they should be exposed somewhere in the region, yet there are no conglomerates mapped in the surrounding 200 square miles (Can. Geol. Surv. Map No. 828). That the conglomerates would have to be of great thickness to furnish so many bowlders is shown by the known extent of the dike: 1,000 feet in length and 900 feet in depth.

2. If the dike had procured its inclusions from a Tertiary conglomerate, now largely eroded, there appears to be no good reason why a large majority of the inclusions should consist of a dynam-

[1] This information was kindly furnished by Dr. Charles W. Drysdale, of the Canadian Geological Survey.

ically metamorphosed rock which is not exposed at the surface. In the Boundary district there occur Tertiary tuffs and conglomerates overlain by basalts and other lavas, the bowlders as well as the matrix of these tuffs and conglomerates being largely of igneous origin. The conglomerates were apparently only locally developed, their formation being preceded and succeeded by volcanic activity.[1] The relative age of the dikes and the Tertiary sediments is not known.

3. The third theory involves shattering along the walls of the dike, the fragments being rounded during their ascent in the dike-rock. The explanation as to why the inclusions (so far as the writer is able to judge from the specimens collected) consist largely of anorthosite must lie in an unusual amount of shattering accompanying the formation of the dike-fissure in that very compact rock, while little shattering occurred in the younger intrusives which may still have been at a high temperature. It is certain that the presence of several million inclusions under the conditions described calls for extraordinary conditions.

The material which has been removed from the inclusions appears in part in the hand specimen of the kersantite as white xenocrysts of varying size. In the field these xenocrysts are not noticed, on account of the weathered condition of the rock. It is also probable that many of the feldspar phenocrysts—and perhaps even some of the pyroxene phenocrysts—of the kersantite are really xenocrysts of the anorthosite.

Little Belt Mountains, Montana: In the Little Belt Mountains Pirrson has described some minette dikes which have brought up masses of a plutonic rock from some depth below the surface.[2] One of the inclusions is a mica syenite, 2 to 3 feet in diameter. The minette does not show the least amount of endomorphic modification, but retains its normal minerals and structure to the contact. Pirrson writes: "From this we may infer that the mass was taken up while the magma was extremely hot, and that it had acquired very nearly the temperature of the fluid mass before the latter

[1] R. W. Brock, "Preliminary Report on the Rossland District," *Can. Geol. Surv.* (1906).

[2] *U.S. Geol. Surv., 20th Ann. Rept.*, Part 3, p. 536.

began crystallizing." The syenite has been somewhat altered by mineralizing vapors from the magma.

The sapphire-bearing Yogo dike also contains inclusions.[1] This mica-trap dike cuts Madison limestone. It is from 3 to 6 feet wide and is vertical. The dike walls are rough, but not especially irregular, and have been slightly indurated by the intrusion. In some places the upward termination is seen, as shown in Fig. 3. The fragments are angular and consist of limestone and shale from the Madison and underlying terranes. In the main excavations a similar breccia is shown at the surface, but the size and number of the fragments decrease with depth.

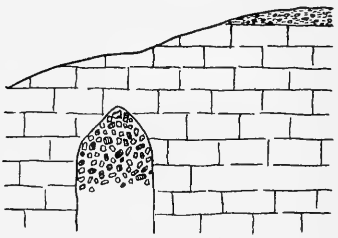

FIG. 3.—Section of upper limit of sapphire-bearing dike, wall of Yogo Canyon, Montana. The inclusions are limestone and shale fragments from the fissure walls, which have floated up to the top of the dike. (After Weed.)

The fragmental material has evidently been shattered from the fissure walls and floated upward as the molten rock rose in the fissure. The Yogo dike is the only case presented in this paper where the inclusions are proven to have floated up to the top of the dike. From the diminution in the number of inclusions with depth, it is evident that if the dike were exposed only at a lower level, few if any inclusions would be shown. On the other hand, inclusions are rarely present at the upward termination of dikes, or, if present, they have not been mentioned in the descriptions.[2]

Syracuse, New York: A few inclusion-bearing peridotite dikes at Syracuse, New York, have been described by several writers. The country rock is of Salina (Silurian) age. The inclusions are largely of Paleozoic rocks, but some are of pre-Cambrian gneisses. The latter are usually more rounded than the former, the result of the attrition involved in their upward journey.[3] In one dike

[1] W. H. Weed, *U.S. Geol. Surv., 20th Ann. Rept.*, Part 3, p. 455.

[2] Suess in *Das Antlitz der Erde*, Vol. III, Part 2, p. 658, describes several dikes the tops of which are exposed, but the above is the only one in which inclusions are mentioned.

[3] C. H. Smyth, Jr., *Amer. Jour. Sci.* (4), XIV (1902), 26.

which is 6 feet in width there are long, narrow inclusions of the country rock, one of which is 15 feet in length and about 1 foot wide with tapering ends. The thickness of the sedimentary rocks in the region is probably about 4,000 feet, and through this distance the pre-Cambrian inclusions have come.

Near Ithaca there are about 25 peridotite dikes, some of which contain a large number of shale and limestone inclusions from the country rock (of Upper Devonian age) and from the underlying Paleozoic sediments.[1] Similarly in the peridotite dikes of Elliott County, Kentucky, there are inclusions of acid (pre-Cambrian?) rocks.[2]

Crazy Mountains, Montana: On the south side of Cottonwood Creek in the Crazy Mountains, Montana, are two elliptical stock-like masses of ouachitite breccia cutting flat-lying Fort Union, Eocene, shales.[3] The masses are 750 feet apart. The lower one is 200 feet long and 40 feet wide, the upper one 300 feet long and 200 feet wide. In each case the contacts are vertical and the igneous masses are jointed in both a horizontal and a vertical direction, the joints cutting both inclusions and cement. The fragments consist largely of fine-grained granitic gneiss, feldspathic quartzite, black slate, and granite. The outline of the fragments is subangular to angular. The interesting feature of these breccias is that the gneissic granite must be of pre-Cambrian age, as no such rock is exposed in the petrographical province. The thickness of the sedimentary series underlying the Crazy Mountains[4] is 24,000 feet, consisting of 4,500 feet of Algonkian, 3,900 feet of Paleozoic, 100 feet of Juratrias, and 15,500 feet of Cretaceous and Eocene strata. Therefore, the gneiss inclusions have apparently risen vertically for a distance of over 4 miles.

The origin of the inclusions was probably by thermal combined with mechanical shattering, as the magma has ascended in elliptical intrusions, replacing the country rock. The circulation in the

[1] *U.S. Geol. Surv., Geol. Atlas,* Folio No. 169, 1909.

[2] J. S. Diller, *U.S. Geol. Surv., Bull. 38,* 1887.

[3] The writer is indebted to Professor J. E. Wolff for the description of these breccias.

[4] W. H. Weed, *U.S. Geol. Surv., Geol. Atlas,* Little Belt Mountains Folio No. 56.

magma must have been very active to bring up blocks of gneiss from such a depth.

Beemersville, New Jersey: Near Beemersville and Libertyville, New Jersey, are several masses of breccia whose form suggests narrow volcanic necks or stocks. The occurrences have been described by Kemp[1] and by Wolff.[2]

The fragments are of various sizes from bowlders to fine grains and consist of gneiss, granite, limestone, and shale (the country rock), imbedded in an ouachitite. The shale inclusions show very sharp angular edges. The presence of inclusions of gneiss show that there has been vertical transportation of nearly a mile. Here again the elliptical-shaped intrusions have probably come up by replacing the rock through which they have passed, and the inclusions probably represent the stoped blocks.

There are also a number of lamprophyric dikes in the region which contain abundant inclusions of the rocks which they cut.

SUMMARY

Inclusions in dikes are rare and are due to special or to accidental causes. The majority of the inclusions have been shattered from the walls of the dike either in the formation of the fissure through which the dike came or in the injection of the dike-magma. The shattering is largely a mechanical operation, but thermal and mechanical action may be combined in many cases. In a few instances the inclusions are pebbles derived from a conglomerate.

The direction and amount of the movement of the inclusions in the several examples are shown in the table on p. 181.

In the cases where some inclusions rise and others sink in the same dike, as at LaTrappe near Montreal, the movement does not appear to depend as much upon the specific gravity of the magma relative to that of the inclusions as upon the mechanics of intrusion and the circulation in the dike-magma before its consolidation. Sandstone fragments have descended in this dike although their specific gravity when in the magma was probably

[1] "On Certain Porphyritic Bosses in Northern New Jersey," *Amer. Jour. Sci.* (3), XXXVIII (1889), 130–34.

[2] *U.S. Geol. Surv., Geol. Atlas*, Franklin Furnace Folio No. 161, p. 13.

less than that of the magma. In the other cases where the inclusions have sunk, the fragments are composed of various rocks, including quartzite, sandstone, slate, limestone, gneiss, and diabase.

Examples	Movement of Inclusions		
	None	Up	Down
Cornwall	×		
Mexico	×		
Cape Ann	×		
Montreal	×	3,000 ft.	2,000+ft.
Marblehead			1,000+ft.
Southern Sweden			×
Cripple Creek			×
Pequawket			×
Brazil		?	×
Shelburne Point	×	12,000+ft	
Mancos		×	
Aschaffenburg		×	
Somerville		×	
Rossland		×	
Ogunquit		×	
Little Belt Mountains		×	
Syracuse	×	4,000 ft.	
Beemersville	×	×	
Crazy Mountains	×	24,000 ft.	

Inclusions in dikes are in general rounded when they have come from some depth and angular when near the place of origin. This indicates that the rounding is largely due to friction which varies with: (1) the distance through which the inclusion has moved, (2) the number of the inclusions; (3) the width of the dike, (4) the temperature of the inclusion at the time of the shattering, (5) the rate of movement of the inclusion.

The initial temperature of the outside of an inclusion is the mean between that of the magma and that of the center of the inclusion. Hence, if there is a fused contact around an inclusion or if the latter shows resorption, a high temperature of the whole fragment is indicated. As most dikes are quickly chilled, it is seldom possible to heat cold inclusions to a temperature at which fusion can take place around their peripheries. In dikes several hundred feet wide, as those of southern Sweden, cold inclusions may be heated to such a high temperature that partial resorption results. In narrower dikes, partial resorption of fragments is possible only

where a special mechanism for the transfer of heat from depth exists for a relatively long period of time, as in a part of the Ogunquit dike.

The material worn from fragments during their rounding and the cement which is dissolved from conglomerates appear in the dike as xenocrysts. Thus, in the hand-specimen of the Rossland dike many small fragments of quartzite and gneiss one-quarter of an inch or less in length may be seen. In the Brevik (Sweden) dike Hedström reports grains of quartz and feldspar of microscopic size derived from the matrix of the conglomerate. In the Somerville dikes numerous xenocrysts of quartz and andesine feldspar, which have come in part from the inclusions, are found in the groundmass. The Shelburne Point dike described by Hitchcock contains many small fragments of the inclusion-rocks.

The question of the fragmentation of rocks from the differential expansion of the constituent minerals and the importance of the inversion point of quartz are of bearing in quartzose inclusions only in the cases where the inclusions have been heated to about 575° C. The criteria of high temperature of the inclusions, at the time of the chilling of the dike, are fused contacts or evidences of corrosion. In part of the Ogunquit dike there is evidence of high temperature and partial assimilation of the inclusions, most of which contain only a small amount of quartz. This corrosion apparently proceeds by the fragmentation of the minerals, as xenocrysts are exceedingly abundant in the dike-rock. In the Somerville dikes the inclusions of feldspar and vein quartz show marked corrosion, yet there is no sign of fragmentation at their peripheries. In the Aschaffenburg dikes there has been a selective formation of xenocrysts of certain minerals as if the rocks had been disrupted by the unequal expansion of the different minerals.

In conclusion it may be stated that some knowledge concerning the mechanism of dike-injection may be gained by the consideration of the movements of the inclusions. Further observations are needed concerning the possibility of circulation which will cause inclusions of lesser specific gravity than the dike-magma to descend in dikes.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA FOR 1909, 1910, 1911, AND PART OF 1912

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II. EASTERN PART OF NORTH AMERICA

Miss Bascom[1] states that the oldest rocks of the Philadelphia district are the Baltimore gneiss and the Wissahickon mica gneiss. Their relation to each other is uncertain. Both are intruded by granites and gabbros.

The Baltimore gneiss presents a granitic and a gneissic phase. The latter consists of alternating, fine-grained layers of mica, quartz, and feldspar. The granitic variety contains quartz, feldspar, and accessories of biotite, hornblende, and garnets. The alternation of layers in the gneiss is regarded by Miss Bascom as evidence of sedimentary origin.

Miss Bascom believes that the Baltimore gneiss is approximately of the same age as the Fordham gneiss of New York, the Stamford gneiss of New England, and the Carolina gneiss of Virginia.

The pre-Cambrian rocks of the Trenton, N.J., district as described by Miss Bascom[2] consist of Baltimore gneiss and the Wissahickon mica gneiss, both of which she regards as of sedimentary origin. Besides these, there are granitic and gabbroic instrusives.

Bastin[3] states that the Adirondack graphite deposits occur in schists of sedimentary origin. Two types of occurrence are recognized, one related to the dynamic metamorphism of carbonaceous

[1] *Philadelphia Folio,* U.S. Geological Survey Folio 162, 23 pp., 12 pls., sections, maps, illustrations, 1909.

[2] F. Bascom *et al., Trenton Folio,* U.S.G.S. Folio 167, 24 pp., 4 pls., 3 figs., 1909.

[3] Edson S. Bastin, "Origin of Certain Adirondack Graphite Deposits," *Econ. Geol.,* V, No. 2 (1910), 134–57.

sediments of the sand clay type, the other to both dynamic and contact metamorphism.

Bayley[1] states that the pre-Cambrian of New Jersey consists essentially of three series of gneisses, sodic, potassic, and basic, respectively, accompanied by pegmatites, all intrusive into still older limestones, the Franklin limestone. The gneisses contain magnetite ores in form pod shaped, northeastward-pitching shoots having the same structural attitude as the gneisses, and are frequently offset by cross and longitudinal faults. The ores are titaniferous, and are associated with basic and alkaline silicates. The Franklin limestone also contains some magnetite ores. These are non-titaniferous, veinlike masses, associated with heavy lime silicates and calcite.

Cushing[2] *et al.* state that the pre-Cambrian succession of the Thousand Islands' region may be classified, beginning with the oldest, as: (1) Grenville metamorphosed sediments, consisting of marbles, schists, gneisses, quartzites which originally were sandstones, shales, and limestones; (2) granite gneisses intrusive into the Grenville, and which have changed the Grenville into amphibolites along the contacts; (3) small bodies of massive syenite, gabbro, and granite; (4) unmetamorphosed basic dikes.

This region shows less complexity of folding, a greater dominance of acid intrusives, and fewer basic igneous rocks than the northern Adirondacks. It is regarded as a transition between the Adirondacks and the Grenville area of Canada.

Dale and Gregory[3] state that the Becket granite gneiss of probably pre-Cambrian age outcrops over wide areas in northern Litchfield county, Connecticut.

Emmons and Laney[4] state that the pre-Cambrian in the Ducktown district consists of the Carolina gneiss, a series of gneisses

[1] W. S. Bayley, "Iron Mines and Mining in New Jersey," *Geol. Survey of New Jersey*, Vol. VII, 499 pp., 23 pls., 2 maps, 31 figs.

[2] H. P. Cushing, H. L. Fairchild, R. Ruedmann, and C. H. Smith, Jr., "Geology of the Thousand Islands' Region," *New York State Mus. Bull.* 145, 194 pp., 63 pls., 6 maps, 9 figs., 1910.

[3] T. Nelson Dale and Herbert E. Gregory, "The Granites of Connecticut," *Bull. 484, U.S. Geological Survey*, 1911. Several maps.

[4] W. H. Emmons and F. B. Laney, "Preliminary Report on the Mineral Deposits of Ducktown, Tennessee," *Bull. 470, U.S. Geological Survey*, 1910.

which probably were developed from the metamorphism of granites, and other igneous rocks and possibly some sediments. It is associated with and probably intruded by the Roan gneiss, which seems to have been diorite and gabbro mainly. Both the Carolina and Roan gneiss are intruded by a younger, less altered granite.

Gordon[1] states that one of the major stratigraphic problems of eastern North America is the separation of the pre-Cambrian from the early Paleozoic sediments and the subdivision of the pre-Cambrian. The problem is complicated by structural intricacy and metamorphism.

Gordon[2] reports that granitic hornblende and mica gneisses, quartzite, and basic eruptives of pre-Cambrian age are found within the Poughkeepsie quadrangle.

Kemp and Ruedemann[3] state that the pre-Cambrian rocks of the Elizabethtown and Port Henry quadrangles of the northwestern part of the Adirondack region may be classified in order of age from latest to the oldest as follows: (1) the unmetamorphosed basaltic dikes; (2) the eruptive complex of more or less metamorphosed granite, anorthosite, syenite, gabbros, and intermediate types; (3) the Grenville series of limestones, ophicalcites, schists, and sedimentary gneisses. The syenites contain lens or podlike bodies of non-titaniferous magnetite with apatite. The hanging wall is generally more acid than the footwall. Locally the gabbros consist dominantly of magnetite and ilmenite.

Kemp[4] summarizes the characteristics of the pre-Cambrian of New York and points out some of its striking similarities to the pre-Cambrian of Sweden.

The oldest pre-Cambrian rocks of the Adirondack are the Grenville sedimentary gneisses and schists, intruded by batholiths of granite gneisses. These were intensely folded and then intruded in the order of time by anorthosite, syenite, basic gabbros, and

[1] C. E. Gordon, "Some Geologic Problems," *Science*, N.S., XXIX (1909), 901–3.

[2] C. E. Gordon, "Progress Report on Poughkeepsie Quadrangle," *New York State Museum Bull. 140*, 1910, pp. 16–20.

[3] "Geology of the Elizabethtown and Port Henry Quadrangles," *New York State Museum Bull. 138*, 1910, 173 pp., 21 pls., 36 figs.

[4] J. F. Kemp, "Pre-Cambrian Formations in the State of New York," Congrès Géologie International, 1910, pp. 699–717.

basalt and porphyry dikes. At Mineville, the syenites are associated with magnetite ores, which in their phosphorus content and in the composition of their wall rocks resemble those of Kiruna. The intrusive episode was followed by intense folding and long-continued erosion. The Paleozoic rocks like those of Sweden were here deposited upon an undulating floor of moderate relief. In New York, evidences of old pre-Cambrian valleys filled by Paleozoics have been found. The visible contacts between the pre-Cambrian and the Paleozoic rocks in the Adirondacks as in Sweden are partly depositional and partly along faults.

The oldest pre-Cambrian rocks in the southeastern area of New York are the Fordham gneiss, consisting dominantly of sedimentary rocks intruded by granite batholiths. Kemp correlates the Fordham gneiss with the Grenville series on the basis of lithologic similarity, and identity of stratigraphic position. Above the Fordham gneiss lie the Manhattan schists and Inwood marble, both metamorphosed sedimentary formations, but less metamorphosed than the Fordham gneiss. They are intruded by plutonic rocks showing a wide range of composition which show parallelism to the Subjotnian of Sweden in their stratigraphic relations, but not in their petrographic characteristics. The Manhattan schists and Inwood limestones Kemp believes may be parallel to the Huronian of the Lake Superior region and the Jatulian of Sweden and Finland.

Kümmel[1] states that the pre-Cambrian rocks comprise a series of basic gneisses and limestones, which are cut by acid gneisses, and pegmatites, all of which lie unconformably beneath the Cambrian. The gneisses are nearly all mineralogically and chemically equivalent to basic and acid igneous rock types, and are largely, if not entirely of igneous origin. The acid intrusives present no evidence of crushing, which probably means that their foliation was developed during crystallization by deformative stresses. The limestones are associated with magnetite ores, and with the Franklin Furnace and Sterling Hill zinc ores. The present state of crystallization, and structural character of the rocks, as well as the develop-

[1] Henry B. Kümmel, "Geological Section of New Jersey," *Jour. Geol.*, XVII, No. 4 (1909), 351–80.

ment of ores, probably resulted from one brief period of deformation and intrusion.

Koeberlein[1] describes the Brewster district as a part of the pre-Cambrian area of New York City, about 54 miles to the northwest of the city. The abandoned Tilly Foster, Brewster, and Croton mines are located here. The stratigraphy from the bottom is reported as Fordham gneiss comprising three series, sedimentary, granitic, and syenitic, respectively, followed by the Inwood limestone and the Manhattan schists intruded by diorites and pegmatite.

The ores are titaniferous and non-titaniferous magnetites, closely related to the syenite of the Fordham gneiss. The syenite contains magnetite, apparently the last mineral to crystallize. The titaniferous magnetites are regarded as magmatic segregations. The non-titaniferous ore of the Tilly Foster mine is regarded as a replacement of limestone by solutions given off during the cooling of the syenite. The gangue consists of chondrodite, garnets, and other minerals characteristic of the deep-seated metamorphism of limestones.

W. J. Miller[2] states that the oldest rocks of the Remsen and Port Leyden Quadrangles of the southwestern part of the Adirondack Mountains are the Grenville sediments, originally sandstones, shales, and limestones which were altered to gneisses and crystalline limestones by syenite intrusions and complex folding, before Cambrian time.

Stose[3] states that the pre-Cambrian rocks of the Mercersburg-Chambersburg area of Pennsylvania consist of a series of altered basalt flows overlaid by altered, finely laminated, spherulitic, and porphyritic rhyolite lava.

Watson[4] reports that an insignificant tonnage of manganese ore has been mined in the crystalline schists of Georgia, which are at

[1] F. R. Koeberlein, "Brewster Iron-bearing District of New York," *Econ. Geol.*, IV (1909), 714–54.

[2] W. J. Miller, "Geology of the Remsen Quadrangle," *New York State Museum Bull. 126*, 1909, 51 pp., 11 pls., 4 figs., 1 geol. map; "Geology of the Port Leyden Quadrangle," *New York State Museum Bull. 135*, 1910, 61 pp., 5 figs.

[3] G. W. Stose, *Mercersburg-Chambersburg Folio*, U.S.G.S. Folio 170, 19 pp., 8 pls., 5 figs., 1909.

[4] Thomas L. Watson, "The Manganese Ore Deposits of Georgia," *Econ. Geol.*, IV (1909), 46-55.

least partly of pre-Cambrian age. Most of the ores occur in a residual clay derived from the decay of crystalline schists. The minerals composing the ores are generally oxides, but silicates are also represented, the latter being somewhat abundant in the undecayed rocks.

Woodman[1] states that the pre-Cambrian mountain protaxis of Cape Breton, Nova Scotia, is not well known. Certain iron-bearing dolomites called the George River limestone may be Upper pre-Cambrian.

[1] J. E. Woodman, "Report on the Iron Ore Deposits of Nova Scotia," *Canada Dept. of Mines*, 1909.

[*To be continued*]

REVIEWS

Atlas photographique des formes du relief terrestre. Prepared by an International Commission *a quarto.* Genève (Suisse): Fred. Boissonas et Cie.

Part I of the first series of plates in this atlas has been issued and distributed. In this part there are eight excellent plates chosen to illustrate the land forms produced by rock disintegration and the action of gravity. These plates were selected from a vast number of subjects submitted by members of the International Commission appointed by the ninth International Congress of Geographers. This commission was directed to co-operate in the preparation of the most comprehensive atlas of relief forms of the earth that has ever been prepared for publication.

The burden of the work has fallen upon the three members of the executive committee, Professors Jean Brunhes, Emm. de Martonne, and Emile Chaix, and to these men much appreciation and praise is due for the excellency and attractiveness of this first number.

The plates included in this first part are the following:

Plate 1.—Disintegration controlled by structure (photographs by M. Lugeon). Disintegration partly independent of structure (photographs by Jean Brunhes, E. de Cholnoky).

Plate 2.—Exfoliation of granite (photographs by G. K. Gilbert, H. W. Turner).

Plate 3.—Exfoliation of granite (photographs by H. W. Turner, C. D. Walcott).

Plate 4.—Peaks formed by frost weathering (photographs by L. Duparc).

Plate 5.—Rock pinnacles (photographs by L. Duparc, E. Chaix).

Plate 6.—Forms due to weathering in the "Dolomites" (photographs by O. Lehmann).

Plate 7.—Crumbling ridge in high mountain and in the polar regions (photographs by E. Chaix, E. de Cholnoky).

Plate 8.—Mechanical disintegration and chemical decomposition combined (photographs by E. Chaix).

Each plate is a phototype printed on special paper 10×13.3 inches. It is accompanied by a description published in three languages, one of which, in each case, is English. These plates and their descriptions

are issued on separate sheets, so that they may easily be made of immediate and permanent educational value in the laboratories of geography and geology, and each such laboratory in America should make arrangements for at least one complete copy of the atlas before the edition is exhausted.

The general plan of the atlas includes:

1. Forms produced by disintegration and the action of gravity: mechanical and chemical disintegration; rock-waste, landslides, etc.

2. Elementary forms produced by erosion by running water: ravining, erosion in swirls, torrents, etc.

3. Complex forms produced by erosion by running water: gorges, valleys; maturity more or less advanced; successive cycles.

4. Forms affected by the character of the rocks: massive, slaty, incoherent, permeable, and soluble rocks.

5. Forms produced by erosion of rocks of various structures: surface features associated with folding and faulting; epirogenic movements.

6. Forms connected with glacial action: existing glaciers, erosion and deposit; ice age.

7. Forms of desert regions: wind erosion; dunes, etc.; complex desert forms.

8. Coastal forms: simple forms due to erosion and accumulation; changes of shore line.

9. Volcanic land forms: accumulation (cones, lava flows, etc.), forms produced by denudation.

The completed atlas will contain 10 series of 6 parts each. Each series will contain from 45–48 plates, so that the atlas will include 450–80 plates and from 1800 to 1900 pages of descriptions, illustrated with maps and diagrams.

W. W. A.

The Fuels Used in Texas. By William B. Phillips and S. H. Worrell. Bull. of the University of Texas, No. 307. 1913. Pp. 287, pls. 22.

The fuels used in Texas are natural gas, oil, sub-bituminous coal, and lignite. All these fuels are produced in the state. For the year 1912 the production of oil was 12,000,000 barrels, 1,200,000 tons of sub-bituminous coal, and 990,000 tons of lignite.

The sub-bituminous coal is mined in three fields. In the north-central part of the state Carboniferous beds carry coal, and on the Rio Grande in two localities coal is worked in beds that are probably Cretaceous. The coal seams are not more than two feet thick in any place.

The object of the report is to emphasize the fact that lignite properly utilized is a valuable commercial fuel. Lignite is well distributed over the state, but it is especially valuable in the northeastern part. It is included in the Timber Belt, Yegua, and Fayette formations of the Eocene period, in the first of which it is most abundant, and there the individual seams reach a thickness of 15 feet.

The value of lignite is shown by chemical analyses and by results of boiler tests. Exclusive of expense of handling, the cost of evaporating 1,000 pounds of water is, on the average, with coal as fuel, 30 cents per pound of water, with lignite, 20 cents, and with oil, 15 cents. Natural gas is the cheapest fuel, but it is obtainable only locally. Lignite may be used as raw coal, as fuel under stationary boilers, as a fuel in gas-producers, for briquetting raw or dried, and for dry distillation and recovery of by-products. It is actually used mostly as raw fuel under stationary boilers. The authors strongly support the report of Dean Babcock of the University of North Dakota that the chief value of lignite will be obtained from its briquetting. They agree that it is almost hopeless to attempt to make useful briquettes out of raw lignite, but that by carbonization (driving-off of the volatile gases), or partial carbonization of the coal and briquetting of the carbon-high product, a commercial fuel may be made which will be of the same order of efficiency as anthracite.

T. T. Q.

Geology of the Titanium and Apatite Deposits of Virginia. By THOMAS LEONARD WATSON and STEPHEN TABER. Virginia Geol. Sur. Bull. No. III–A. 1913. Pp. 300, pls. 37, figs. 22, map 1.

The titanium and apatite deposits of Virginia are in the northwestern part of the state. Rutile is found both in massive syenite and in nelsonite dikes. The apatite is most abundant in some of the nelsonite dikes. The report discusses the subject of titanium quite fully. The chemistry of titanium, its ores, their distribution, and the uses of titanium products are treated fully. Titanium is used in metallic alloys, such as ferro-titanium, cupro-titanium, etc., for lighting purposes as incandescent media, as mordants, as refractory coloring materials for the ceramic industry, etc.

The rocks in the area are igneous in origin and distinctly metamorphosed. They form a peculiar comagmatic group which is characterized by the abundance of ilmenite, rutile, and apatite. These minerals

are the economic values. Other peculiar mineral constituents are blue, opalescent quartz, hypersthene, and hornblende derived from pyroxene. Named in order of their probable differentiation, the principal rock types of the area are: biotite-quartz, monzonite-gneiss, syenite formerly referred to as pegmatite, gabbro, nelsonite, and diabase.

The geology of the ore deposits is fully discussed with especial description of the unusual rock types. The ore minerals occur most abundantly in the nelsonite. The chief constituents of this rock are apatite and ilmenite; there are varieties of nelsonite which are rich in rutile. The nelsonite bodies are of remarkably uniform granularity, irrespective of the size or of the width of the body. Some show evidence of mineralogical gradation into the composition of the wall rock, whereas others show sharp differentiation of the nelsonite and the inclosing rock. It is supposed that the nelsonite dikes were intruded into the still very hot, and possibly in some places unsolidified, rock. The rutile-rich bodies appear to have been formed in advance of the ilmenite nelsonites; this is suggested by their irregularity in boundary, and by their appearance of transition into the surrounding rock like segregations rather than intrusions. Both types are regarded as differentiation facies of the inclosing rocks.

This district contains the richest and greatest known deposit of titanium minerals in the world. With an increase in the utilization of titanium products it is expected that this will become a productive mining region. The apatite deposits will not be of much importance commercially while Florida remains as productive of phosphates as at present. The Virginia deposits both of titanium and of apatite have been worked intermittently but the present output is insignificant.

T. T. Q.

The Ore Deposits of Northeastern Washington. By HOWLAND BANCROFT. U.S. Geol. Surv. Bull. No. 550. Pp. 215, figs. 26, pls. 19.

Except for a brief discussion of the general geologic features of northeastern Washington by way of introduction, this bulletin is given over for the most part to descriptions of mines and prospects in the various districts included. In the discussion of each district some attention is given to the origin of the ores, but the discussion is broadly general. A section on the Republic District, written by Lindgren, is included in the volume.

A. D. B.

VOLUME XXIII NUMBER 3

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APRIL-MAY 1915

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NOTE.—Through oversight, the captions for figures were omitted from the first part of this article, which appeared in the preceding number of this *Journal* (pp. 97–117). These captions are given below:

FIG. 1.—General map of the district

FIG. 2.—Map showing routes traveled by the writer and companions in the summers of 1910–13.

FIG. 3.—Model representing the topography and structure of Wind River basin and vicinity as it probably existed just before the close of the Cretaceous period.

FIG. 4.—Model of folds and overthrusts produced at the close of the Cretaceous, restored approximately on the Carboniferous formations. The actual surface produced by folding even if not eroded would doubtless be of a different nature. The region included extends from the Idaho state boundary east to the Bighorn district. (The flat space in the southwest is merely blank.)

FIG. 5.—Sketch of the surface at the base of the Tertiary sediments north of Circle, as it probably appeared in early Eocene time. (Adapted from a photograph of the modern surface.)

FIG. 6.—Pinyon conglomerate containing irregular sandstone lenses. Hackamore Creek in the Mount Leidy highlands.

FIG. 7.—Stereogram of part of the Wind River basin and vicinity as they may have appeared after the deposition of the Eocene and Oligocene sediments.

FIG. 8.—Folded Eocene beds at Indian Meadows, near Circle. All the strata exposed belong to the typical "Wind River" clays. Drawing from a photograph.

FIG. 9.—Photograph and diagram of the fault near the mouth of Dinwoody Creek.

FIG. 10.—Diagram of the apparent relations of the Eocene conglomerate to the Paleozoic along the south side of the Gros Ventre Range.

FIG. 11.—Stereogram of the Wind River Range and vicinity as they may have appeared (neglecting erosion) after the warping and faulting, probably of Miocene date. Cf. Fig. 7.

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THE

JOURNAL OF GEOLOGY

APRIL-MAY 1915

POST-CRETACEOUS HISTORY OF THE MOUNTAINS OF CENTRAL WESTERN WYOMING

ELIOT BLACKWELDER
University of Wisconsin

PART II

Later Tertiary planation.—Whether or not the early Tertiary formations ever covered the entire district, it is evident that they have been stripped from large areas which they formerly occupied; for scattered outliers are now situated several miles beyond the edge of the continuous outcrop. It is also an obvious fact that large quantities of the original Tertiary strata have been carried away by erosion in the broad basins themselves. The deformation which succeeded the laying-down of the sediments, probably about the middle of the Miocene, must have initiated a new cycle of erosion, and subjected to denudation large tracts which had previously been sites of aggradation. In later pages it will be shown that between the Miocene and the present there have been several erosion cycles, and the interpretation of those cycles forms an important part of this essay. The data for this must be drawn largely from the features of the existing topography, now to be considered.

The central part of the Wind River Range is in reality a broad dissected plateau 20–30 miles wide, surmounted by a narrow axial range of sharp peaks. The plateau is no longer complete

but rather consists of numerous tabular remnants,[1] which the eye can easily combine and thus reconstruct in imagination the original surface (Figs. 12, 13, 14). Such a reconstruction suggests a region reduced largely to flatness, but with subdued mountains and hills with a relief of less than 1,500 feet near the divide, i.e., a state of early old age. The topography was largely independent of rock structure, passing indiscriminately

FIG. 12.—Photograph of Green River lakes and Square Top Mountain, a remnant of the summit peneplain. Photograph by C. L. Baker.

across hard and soft members of the Archean complex and also across the alternately hard and soft Paleozoic strata. In 1909 this was recognized as a peneplain by C. L. Baker;[2] and Westgate and Branson[3] in a later paper described a similar feature at the

[1] See the Fremont Peak, Wyoming, topographic sheet of the U.S. Geological Survey.

[2] C. L. Baker, "Cenozoic History of Western Wyoming," *Bull. Geol. Soc. Am.*, XXIII (1912), 73 (abstract).

[3] L. G. Westgate and E. B. Branson, "Later Cenozoic History of the Wind River Mountains," *Jour. Geol.*, XXI (1913), 142–59.

southeast end of the Wind River Range. Instead of assuming the origin of the plain, it is fitting that we examine the question before proceeding.

It is apparent that the plateau surface is a cut plain, with residual hills and mountains still standing above it. There appear to be three methods by which extensive cut plains may be produced in the interior of a continent: (*a*) by the action of ice-sheets;[1] (*b*) by the long-continued action of the wind; (*c*) by river systems that have endured to old age.

FIG. 13.—Mountain ridge (11,500 feet) near Gypsum Creek at the west end of the Wind River Range, showing a syncline in Paleozoic rocks cut off at the level of the peneplain.

There is so little in favor of the hypothesis that the plain was produced by ice-sheet glaciation, that it may be quickly dismissed. It will be shown that the plain is one of the oldest topographic features in the district and therefore probably pre-Quaternary. It now carries locally upon its surface deposits of residual soil. There is no evidence of general ice-sheet glaciation in the surrounding districts. It would have to be assumed that the present axial range has been developed since the ice sheet disappeared, for ice sheets inevitably tend to destroy rather than to make such features,

[1] F. E. Wright, "Effects of Glacial Action in Iceland," *Bull. Geol. Soc. Am.*, XXI (1910), 717–30.

and would almost surely succeed if they persisted so long as to planate the surroundings.

It seems to be established that prolonged wind abrasion in deserts does produce plains on which only scattered elevations remain; although the criteria for recognizing ancient wind-made plains are hardly as well agreed upon by physiographers as are the criteria for peneplains. Such a plain, if undefaced, should have certain characteristic features which could be used in recognizing its origin. Its depressions should be broad, relatively flat-bottomed

FIG. 14.—Photograph of mountain (11,500 feet) near Gypsum Creek. Tilted Paleozoic limestones trunkated at summit.

hollows, rather than graded valleys, and these hollows should have a definite relation to the weakest rocks. Wherever hard and soft rocks are exposed there should be the characteristic wind-etched ridges, ledges, mushroom rocks, etc. The pebbles in the gravel strewn upon the flats by occasional floods should be pitted, polished, and even beveled by the sand-blast. Only the most massive and resistant bodies of rock should stand out as hills and mountains.

Unfortunately it is almost impossible to apply the test of these criteria to the Wind River plateau in its present condition. The details of its surface forms and material have been greatly modified by the long-continued action of frost above timber-line. The

trenching of the plateau by many canyons, and especially the scouring of its surface by widespread alpine glaciers in a comparatively recent epoch, have generally effaced the original details of its topography. The higher mountain peaks do not, however, appear to have any relation to especially hard rocks, but rather form a definite range or divide near the crest of the anticline. In short, if the plain was largely fashioned by wind action, all evidence of the fact seems to have disappeared.

The competency of streams to reduce hard and soft rocks alike to monotonous relief, without removing the last residual mountains, is generally admitted. Upon such a plain there should be only thin alluvial deposits and widespread residual soil. It is an observed fact that on the southwest side of the Wind River Range

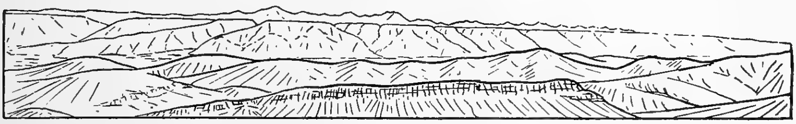

Fig. 15.—Drawing (from a photograph) of Wind River peneplain and range as seen from Triangle Peak in the Gros Ventre Range.

deposits of soil, several feet deep, due to the decay of the granitic rock in place, still exist upon the tops of mountains 11,300 feet high, which constitute a part of this or a still younger plain (Fig. 16). Since the original drainage can no longer be seen, and since in all but a few places topographic detail has been modified by subsequent glaciation, the evidence directly in favor of river erosion is but little more satisfactory than that for wind action. The most significant fact is perhaps the continuous range of peaks, forming a divide independent of rock structure. Such divides are characteristic of stream-made topographies, but of no others. As the problem now stands there is then only a measure of probability, but not proof, that streams were the agents of planation; and so, for convenience, the old plateau surface will be called in later pages a "peneplain," implying thereby only that it was produced by the long-continued action of degrading agencies.

If reconstructed by filling the depressions cut out of it, the summit peneplain would stand forth as a gently undulating

surface, surrounded by many low hills and even mountains of gentle slope, rising to a distinct divide now marked by the crest of the Wind River Range. Unmarred remnants of this surface are believed to exist still in Goat Flat, the high plateau west of West Torrey Creek, the flats both north and south of Clear Creek and other similar features ranging from about 12,000–13,000 feet in altitude, and declining near the outer borders of the range to about 11,300 feet. As a result of two or three stages of glaciation, the slopes of the once broad, residual hills and especially the axial

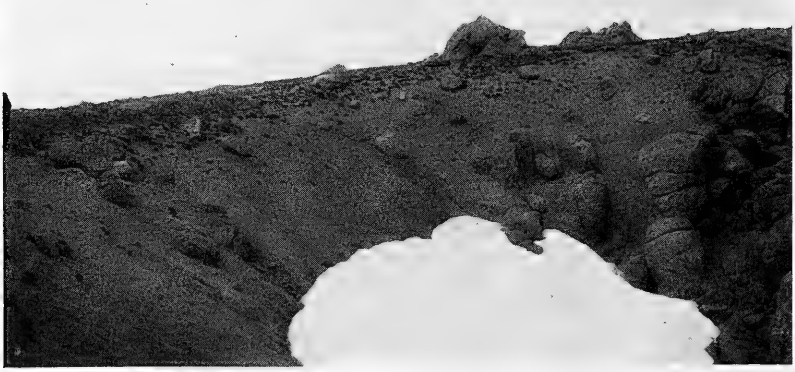

FIG. 16.—Residual soil from granite, a remnant of the Wind River peneplain, now at 11,300 feet, Gypsum Creek.

divide have been excavated and made much steeper than before. To this process I ascribe the existing sharpness of the peaks and arêtes of the range. The broad plateau remnants at 10,000–10,500 feet east of Pinedale may not be parts of the summit peneplain, but rather the result of a later cycle.

Although the ancient peneplain has since been widely demolished by erosion, other remnants of it than the Wind River plateau should exist; but a search of neighboring ranges brings unsatisfactory results. The Mount Leidy highlands, to the northwest, are so maturely dissected that no ancient flats remain at high elevations. The flats at 11,000–12,000 feet in the Absaroka Range

may well belong to the same stage, but they are less significant because they lie upon horizontal strata. The more or less level crest of the Hoback Range, at 9,000–10,000 feet, is somewhat more suggestive of planation. This range consists of highly folded hard and soft rocks, which if planated and then dissected would cause the sculpturing of level-topped ridges. This type of topography prevails southward along the Wyoming Range and is conspicuous near Labarge Mountain northeast of Kemmerer. Along the crest of the Gros Ventre Range and especially in the western half there are some flats of considerable area at elevations of 10,300–10,800 feet. These have been cut across strata of varying hardness tilted at angles of 10–50° on both flanks of an anticline. However, it should be noted that they lie 1,000–2,000 feet lower than the Wind River plateau. As the mountain ridges to the southwest are considerably lower (8,000–9,500 feet), they may belong to later peneplains, unless the Wind River surface has been notably warped.

From the canvass of the district it appears that the peneplain is preserved best on the very massive hard rocks of the broad Archean outcrop in the Wind River Mountains. Somewhat doubtfully I correlate with this the smaller flats along the crests of ranges where hard folds in Paleozoic rocks have been trunkated. Over the intervening territory the weakness of the Mesozoic and Tertiary rocks has enabled the streams to destroy the peneplain.

In the Teton Range there is a broad expanse of the massive Archean formations, and hence the preservation of another large peneplain remnant might be anticipated. Instead we find an extremely rugged surface with but few small elevated flats. The east base of the range is marked by a fault of more than 10,000 feet displacement, along which the soft Cretaceous shales and Tertiary clays have been brought down to the level of the massive Archean gneiss. There is good evidence that this fault is not younger than the mid-Tertiary epoch of diastrophism. Urged by such conditions, the rejuvenated streams are believed to have rapidly excavated Jackson Hole from the Cretaceous and Eocene beds, thus leaving the Archean mass of the Tetons to form a wall of imposing height along the west (Fig. 17). Consequent streams on the face of the wall were able to cut only short steep gorges in the block,

while the Snake River was rapidly excavating the adjacent beds of clay and shale. That portion of the peneplain upon the

FIG. 17.—The eastern front of the Teton Range, including the Grand Teton. (Photograph by Stimson)

Teton block thus found itself in a position particularly vulnerable to the attack of streams and glaciers, and hence it has been carved into ruggedness more rapidly than other uplifts. Nevertheless

at least two small remnants of an apparent peneplain still exist north of the Grand Teton; one is the flattish summit of Mount Moran (12,100 feet) and the other a triangular flat summit at 10,000 feet on the north side of Birch Creek several miles to the west. Only the latter of these has ever been examined by a

Fig. 18*a*.—Waterworn cobbles of quartzite among limestone fragments on the flat summit of a peak (10,100 feet) in the Teton Range.

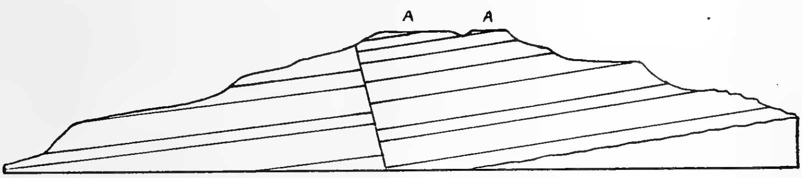

Fig. 18*b*.—Diagram of the occurrence of the gravel shown above. The gravel lies on the surfaces (*AA*).

geologist (Figs. 18*a* and 18*b*). Its surface is strewn with the usual frost-cracked débris of the underlying Madison limestone, but also with well-rounded cobbles of quartzite which still retain their original polish. That this is a bit of an old stream-worn surface is thus clearly shown. It may be a part either of the Wind River or later peneplains, or of the unconformity at the base of the Pinyon conglomerate. It may be argued that the peneplain would not

be likely to have coarse gravel upon its surface. The surface underlying the Eocene formations, which is gravel-strewn, has already been described as being rugged.

In Yellowstone Park the formation of the peneplain may have been prevented, or if it was made, it may have been overwhelmed, by the volcanic eruptions which students of that district assign to the Pliocene and later epochs.

Turning to other parts of the Rocky Mountains, I find elevated peneplains widespread. In the Bighorn Range[1] there seems to be a similar high plateau surmounted by an axial mountain range, which, prior to the cirque-making by Quaternary glaciers, had a subdued post-mature topography. There also the eroded surface trunkates the various rock structures indiscriminately. In another place[2] I have described the peneplain which is well preserved in the Sherman Mountains of southeastern Wyoming. Other cases of somewhat similar characteristics have been reported from Colorado, Montana, Idaho, Washington, and other western states. These peneplains have been assigned to the Eocene, Miocene, and Pliocene,[3] but they are so similar in their characteristics and relations as to suggest that they may be only local remnants of a once widespread old-age surface, subsequently more or less warped.

The age of the Wind River summit peneplain is debatable, but, by a process of finessing, it may be worked out with some degree of assurance. It will be admitted by all that, since it trunkates the structures produced by the folding at the close of the Cretaceous, it must be of Cenozoic age. The stage of topographic old age, when planation is being accomplished, does not permit the accumulation of thick sedimentary deposits in the same district because the two processes are mutually exclusive. This appears to be true whether the work is done by wind or by streams. For that reason it is hardly conceivable that the peneplain could have been developed while the Eocene and Oligocene strata were being deposited to thicknesses of several thousand feet in various parts

[1] See Plates VIII A, XXVII A, XXX, and XXI A, in "Geology of the Bighorn Mountains," by N. H. Darton, *U.S. Geol. Survey, Prof. Paper 51*, 1906.

[2] *Jour. Geol.*, XVII (1909), 429–44.

[3] See p. 194.

of the same region. If this inference is sound, the peneplain must have been made either just before or just after the Eocene-Oligocene epoch of deposition. We may first consider the hypothesis that the peneplain is part of the pre-Wasatch (Lower Eocene) topography.

The eroded surface which underlies the Eocene strata is clearly exposed at many points among the ranges of western Wyoming. Where it trunkates the soft Mesozoic strata, this surface is relatively flat, but on the harder rocks nearer the axes of the ranges, it is hilly and even mountainous in relief. Thus in the southwest part of the Kirwin quadrangle, between Double Diamond ranch and Mountain Meadows, the Wind River Eocene beds rest upon an eroded surface which has a visible relief of more than 1,200 feet. Immediately upon it there generally lies a coarse conglomerate (Pinyon) of variable thickness. Northeast of the Gros Ventre Range this reaches a thickness of about 1,000 feet, and on the south side it is apparently even thicker. From these facts we seem compelled to infer that there were conspicuous hilly or mountainous tracts when the early Wasatch deposits were laid down, and that streams later leveled up the lesser inequalities with their deposits.

As a second test of the hypothesis of Eocene age, attention should be turned to the relation between the peneplain and the Eocene formations, for if those relations are clear they may decide the age of the plain. The Eocene deposits now generally lie at much lower elevations than the existing remnants of the peneplain, and the basal contact in some places is more than 6,000 feet below it. I can conceive of but three ways (Fig. 19) in which this could be brought about: (*a*) the Eocene sediments may represent the filling of Eocene valleys excavated in the peneplain; (*b*) they may be remnants of once horizontal deposits laid down upon the peneplain and now preserved by down-warping or down-faulting; or (*c*) they may be bodies of sediment, either deposited in earlier depressions or warped or faulted down, upon the surface of which, together with adjacent formations, the peneplain has subsequently developed. In the first two cases the age of the peneplain would be pre-Wasatch. In the last case it would be post-Oligocene. These three possibilities will now be examined briefly.

If the Eocene sediments were deposited in basins excavated below the peneplain, the sediments should have been derived in part from the slopes and divides of the basins themselves. On both sides of the Gros Ventre Range, however, the thick coarse conglomerate in the Lower Eocene consists not of local rocks but almost entirely of varicolored quartzite with occasional porphyries. Since there are no rocks of this character entering into the make-up of the Gros Ventre Range, these pebbles must have been imported; and the nearest appropriate outcrops are about 100 miles to the northwest.

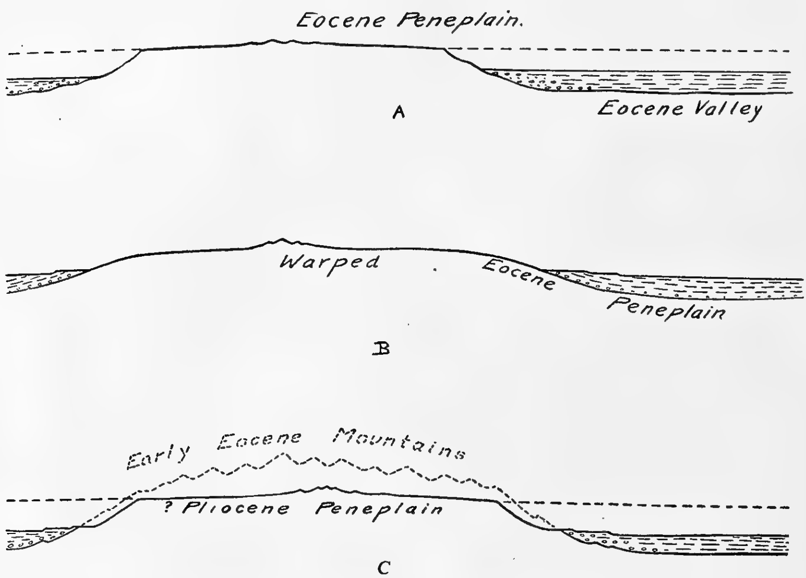

FIG. 19.—Diagrams illustrating various hypotheses as to the relation and age of the peneplain: (*a*) as an Eocene plain with valleys excavated in it and then filled; (*b*) as an Eocene plain covered with sediments and then warped; (*c*) as a Pliocene (?) plain cut across Eocene and older rocks.

Considerations of time are also somewhat adverse, since upon this hypothesis the planation, involving the most massive and resistant rocks, must have been completed in that fraction of the Eocene epoch between the folding of the Paleocene (Fort Union) strata and the deposition of the Lower Eocene (Wasatch) beds. In the same interval of time, spacious depressions must also have been excavated beneath the new peneplain for the reception of the

Eocene sediments. On the other hand, the remnants of the peneplain must have been preserved on a large scale from the early Eocene to the present. Without convincing evidence in its favor this is not, therefore, an attractive hypothesis.

On the supposition that the surface we now find beneath the Eocene sediments is part of the old peneplain, which has merely been warped and partly uncovered in subsequent time, we should anticipate that the softer Paleozoic and especially the Mesozoic strata would be reduced to base-level long before the massive granitoid rocks of the Archean were worn low. Nevertheless, there is abundant evidence that the basal Eocene surface was only maturely hilly upon the outcrops of the Paleozoic and, in some places, of even the Mesozoic strata. That surface was a plain only on the weakest beds. Again, if we are to ascribe the preservation of the Eocene to down-warping after it was deposited upon the peneplain, we should rightly expect to find the Eocene strata parallel to the base on which they rest. On the contrary, in many places they grade horizontally into thick beds of conglomerate which themselves rest against a steeply inclined surface of the older rocks, thus indicating that the beds were actually deposited in a valley or basin and not upon a plain.

To the suggestion that the peneplain is younger than the Eocene sediments, and has beveled them as well as the older rocks, I can find no objection. It is free from the difficulty occasioned by the Miocene deformation of the entire district. It explains the absence of Eocene remnants upon the peneplain and does not conflict with the existence of the rugged basal Eocene contact overlain by horizontal beds. It also corresponds with determinations in other parts of western United States, that the principal peneplains are not older than the Miocene. I therefore accept, for the present, the conclusion that the peneplain is younger than the early Tertiary strata, and later than the mid-Tertiary deformation that expressed itself in warping and faulting.

In subsequent pages it will be shown that a number of events requiring a relatively long time must be referred to the Quaternary period, and there is so much good evidence all over the western states that the opening of that period was characterized by

noteworthy changes of level, that it is probably safe to refer the peneplain to the time between the Middle Miocene and these early Quaternary disturbances.

The complete planation of soft rocks may be accomplished in a geologically short time, but the reduction of very resistant granitoid rocks to a peneplain must require a vastly longer time. Planation of either kind of rocks by streams demands a constancy of conditions which is possible only when the lithosphere is in a state of rest. Inasmuch as such quiet is rarely local only, and if long continued would permit the planation of very large areas, it is worth while to inquire if peneplains were made in adjacent regions, and if so at what time geologically.

Within the past decade several geologists[1] who have independently recognized peneplains in the Rocky Mountains have referred them to late Tertiary time or specifically to the Pliocene. In several other districts somewhat more remote,—the Cascade Range of Washington,[2] the Sierra Nevada in California,[3] and the Grand Canyon of Arizona,[4]—peneplains have been reported and their ages determined as approximately Pliocene. The summit peneplain of central Idaho is assigned by Umpleby[5] to the Eocene, but in a review[6] I have suggested that the opinion is not well founded.

In brief, the evidence from all points of view here considered, although it does not establish the age of the peneplain, does strongly

[1] W. W. Atwood and K. F. Mather, *Jour. Geol.*, XX (1912), 407; S. H. Ball, *U.S. Geol. Survey, Prof. Paper 63*, 1908, 32; E. Blackwelder, "Cenozoic History of the Laramie Region," *Jour. Geol.*, XVII (1909), 437; J. L. Rich, "Physiography of the Bishop Conglomerate, Southwestern Wyoming," *Jour. Geol.*, XVIII (1910), 613 (age given as post-Oligocene, probably Miocene); L. G. Westgate and E. B. Branson, "Cenozoic History of the Wind River Mountains," *Jour. Geol.*, XXI (1913), 144.

[2] B. Willis and G. O. Smith, "A Contribution to the Geology of the Cascade Mountains," *U.S. Geol. Survey, Prof. Paper 19*, 1903, 70.

[3] F. L. Ransome, "The Great Valley of California," Univ. of Calif., *Bull. Dept. of Geol.*, 1896, I, 371–428.

[4] H. H. Robinson, "A New Erosion Cycle in the Grand Canyon District, Arizona," *Jour. Geol.*, XVIII (1910), 742–63.

[5] J. B. Umpleby, "The Old Erosion Surface in Idaho," *Jour. Geol.*, XX (1912), 144.

[6] E. Blackwelder, "The Old Erosion Surface in Idaho: A Criticism." *Jour. Geol.*, XX (1912), 410-14.

indicate that it was Pliocene. At that time western Wyoming must have been in a state of tectonic rest which permitted the streams or the wind, or both, to reduce nearly all of the district to the condition of a peneplain, interspersed with post-mature mountain masses, situated on the most resistant rocks (Fig. 20). Appropriately we find in the region no sediments of presumptive

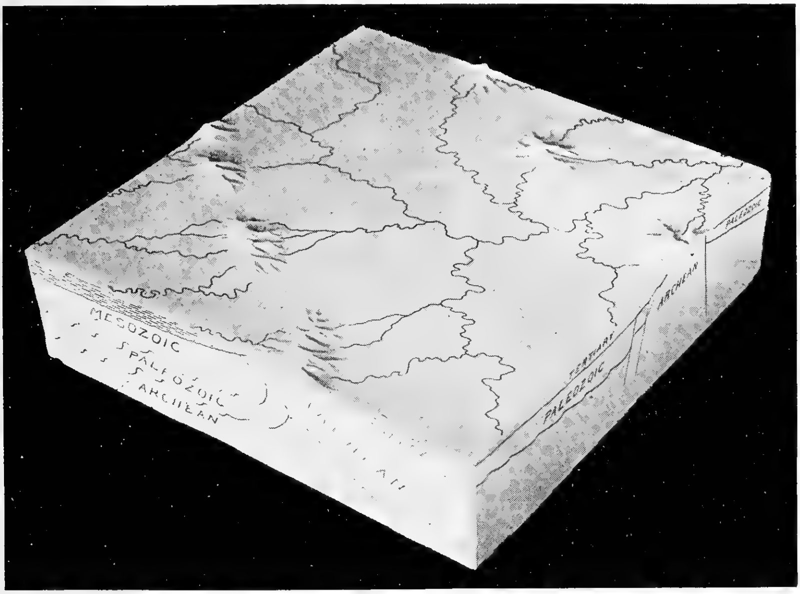

FIG. 20.—Model of a part of the Wind River basin and vicinity as it may have appeared at the completion of the peneplain.

Miocene or Pliocene age and hence know little of the climate, the life, or the changes of the time.

Early Quaternary rejuvenation.—Even where now best preserved, the peneplain is trenched by deep valleys, and over a large part of the district it has been destroyed by their growth (Fig. 21). The remnants now stand at elevations (11,000–13,000 feet) so high that no agency of land sculpture seems capable of fashioning them as they stand. Since the graded streams of the district now flow in channels between 5,000 and 8,000 feet above sea-level, the remaining parts of the peneplain are obviously in a vulnerable

position, and the progressive lowering of grade makes them only more so. The present altitude of the remnants may therefore be ascribed to movements in the lithosphere. Two possibilities suggest themselves: (*a*) local warpings whereby small parts of the peneplain were elevated or their surroundings were depressed; and (*b*) a widespread uniform change of level which incited the erosive agencies to excavate the soft rocks deeply, and so left only remnants of the peneplain standing out in relief, where the rocks were most resistant.

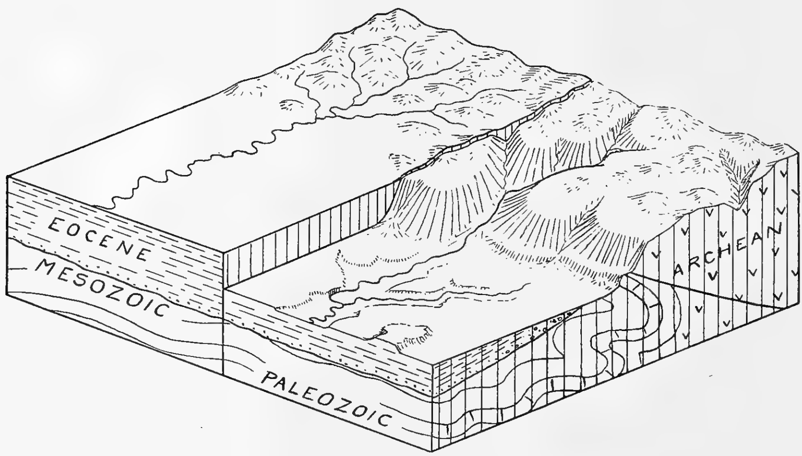

FIG. 21.—Diagram of a part of the Wind River Range before and after the dissection of the peneplain.

If the Wind River plateau be regarded as a local uplift, it may be either a horst or an upwarp (Fig. 22). That it is not the former is sufficiently indicated by the observed fact that it is not surrounded by faults. If it were an upwarp the peripheral slopes would be likely to preserve remnants of the peneplain bent downward but otherwise showing the trunkation of the various structures. I have looked carefully for such features in the foothills of the range, but with negative results. On the contrary, the peneplain comes out to the west front of the range at a nearly constant elevation and breaks off abruptly to the lowlands west of it. These facts suggest that the present contrast between the the isolated plateau and the surrounding plains is not due primarily

to differential warping; but they do not preclude the possibility of warping in a minor degree.

The hypothesis that existing conditions are due to uniform regional uplift followed by selective denudation is the only one

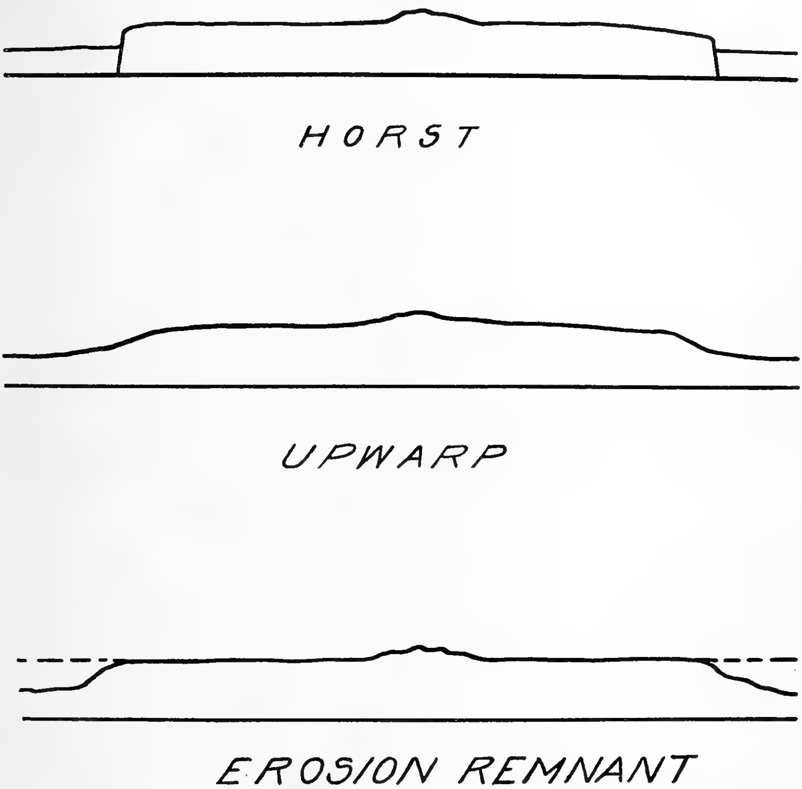

FIG. 22.—Diagrams illustrating the possible origin of the existing Wind River plateau as a horst, an unwarp, or an erosion remnant.

against which I find no serious objections. It agrees well with the fact that remnants of the peneplain now exist only upon the outcrops of the hardest rocks, and with the other fact that these remnants now stand at a more or less common elevation. The hypothesis of regional uplift is therefore accepted tentatively.

Local evidence alone is not sufficient to permit a close determination of the age of this elevatory movement. Obviously it followed the development of the Wind River peneplain, which the available evidence indicates occurred in the late Miocene and Pliocene. In later pages it will appear that there have been several changes of level separated by long intervals of quiet. All, however, seem to have occurred before the last stage of glaciation. It is significant that many of the geologists who have studied the western mountain states have recently concluded that the beginning of Quaternary time was marked by important changes of level, to which the present relief and ruggedness of our western

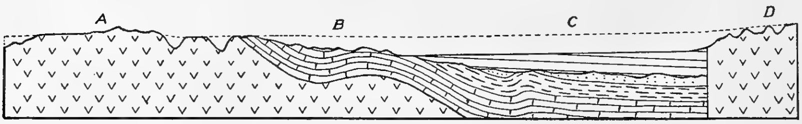

FIG. 23.—Diagram illustrating the progress of the same erosion cycle in rocks of three different degrees of hardness: (*A*) residual areas of older surface with canyons in granitic complex; (*B*) maturely mountainous surface on Paleozoic limestones, etc.; (*C*) plain on Tertiary clays: (*D*) topography of granitic mountains more advanced than at *A* because of proximity to deep basin.

mountains is in large measure due. These considerations lead to the somewhat indefinite conclusion that the several changes of level occurred between the late Pliocene and the close of the Pleistocene epoch.

General effects of erosion in the Quaternary period.—On every hand there are deep canyons, wide river basins, and other speaking evidences of denudation in geologically recent times. It is reasonable to believe that the district has been subject to erosion under slowly varying conditions and hence with varying results ever since the beginning of the uplift which seems to have accompanied the close of the Tertiary period. Although I am not disposed to belittle the efficiency of the wind in fashioning topographic details, nor in deporting dust from the region and in that way gradually lowering its elevation, it is clear from their forms that nearly all the coarser and many of the finer topographic features of western Wyoming have been carved by running water, and that in general the topography of the district has been under the control of the three prin-

cipal water drainage systems—the Snake, the Green, and the Bighorn. Glaciation and eolation have merely modified the effects of erosion—the one locally, the other generally.

Before proceeding to describe the effects of stream denudation in the district, attention is called to the fact that the surface is underlain by a great variety of rocks which differ vastly in the resistance which they offer to erosive processes. Under a subarid

FIG. 24.—Plane surfaces developed upon the Wind River Eocene beds on the flanks of Black Mountain, in the Wind River basin.

climate, the Tertiary sandy clays are the weakest rocks; the massive Archean gneisses and granites stand near the other extreme of resistance. Between them all gradations exist. It is evident that in the softest of these materials, erosion proceeds with great rapidity in contrast with its progress upon the most refractory formations. This fact (in addition to the structure) is the key to the origin of the principal topographic features of the region (Fig. 23).

In the notably weak beds of Tertiary and Cretaceous age, broad plains have been developed at elevations several thousand feet below the surrounding highlands (Fig. 24). Good examples

of these are the Wind River basin, Jackson Hole, Teton basin, Bighorn basin, Green River valley, and many others outside the district here considered. In these wide basins the younger rocks, down to the more or less resistant Triassic(?) red beds, or even down to the Carboniferous terranes, have nearly all been planed off to a common level. It is highly probable, however, that all of

FIG. 25.—Bad-land topography in the Wind River basin. Shows dominance of the effects of running water even in the driest part of the district.

the basins were formerly filled more or less completely with the soft Tertiary strata, and that most of the erosion has been accomplished in them rather than in the older rocks beneath. All are still partly filled with formations of that age, and outliers show a much more complete filling before erosion occurred.

The floors of these basins are by no means simple flats (Fig. 25), but the details will be discussed more fully in later pages.

At the intertwined headwaters of the three great river systems, the valleys are still narrow and in part ungraded. There, even

upon the soft Cretaceous and Tertiary beds, erosion has produced only a maturely dissected surface (Fig. 26). In this district the peneplain has been wholly destroyed, and it is improbable that any of the summits now rise to its former elevation.

In the moderately resistant limestones, sandstones, and interbedded shales, which constitute the Paleozoic sequence, erosion has proceeded much less rapidly than in the soft Mesozoic and Tertiary formations. Thus where the former rise to the surface we find them only maturely dissected into ramifying V-shaped

FIG. 26.—Mature topography developed on weak Cretaceous and Eocene strata along Fish Creek, near the Continental Divide.

valleys separated by skeleton ridges, or, where folded, into parallel mountains and valleys. The topography of the Gros Ventre Range and of the west slope of the Teton Range, as well as of mountains farther south along the state boundary, illustrates this phase of topographic development.

Erosion has had the least effect upon the hard, massive, and tolerably uniform granites and gneisses of the pre-Cambrian complex. In the district under consideration there are but six localities where these ancient rocks reach the surface, and in three they have merely been uncovered in the bottoms of mountain canyons. The only expansive outcrops at high elevations are those in the Teton,

Wind River, and Owl Creek ranges. In the Teton Range, the juxtaposition of very hard and very soft rocks along a fault has created abnormally steep slopes,[1] so that the Archean rocks of the

[1] The magnificent east front of the Teton Range can scarcely fail to suggest to the geologist a recent fault scarp, but the consideration of it from various angles in three different seasons has convinced me that it is in fact a "fault-line scarp" (W. M. Davis, *Bull. Geol. Soc. Am.*, XXIV [1913], 187–216) situated along a Middle Tertiary displacement, of which the original topography has long since been destroyed. The facts which suggest a recent fault scarp are the abruptness and continuity of the wall, the comparative straightness of the base, and the triangular facets at the distal ends of the very short steep spurs. The gravel-strewn floor of Jackson Hole seems also

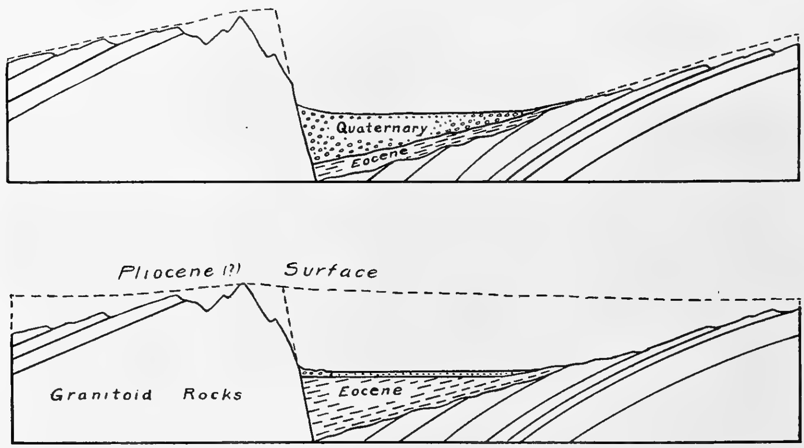

FIG. 27.—Diagrams of the Teton front to explain two possible origins of the escarpment: (*a*) as a modern normal fault scarp; (*b*) as a fault line scarp discovered entirely by erosion.

to indicate the thick alluvial deposits normally made on the downthrown sides of new faults. On further consideration, however, it appears that all of these phenomena may be satisfactorily explained without assuming a recent fault, and there are additional facts which indicate that the scarp was produced by the exhuming of a once-buried fault surface. Such a scarp would be abrupt and continuous, and as straight as the fault trace. In the early stages of dissecting the scarp, triangular facets would be developed at the distal ends of the spurs. The alluvial filling of the Snake River valley is of unknown depth, but the exposures of the underlying rock in several parts of the floor of Jackson Hole suggest that there is only a veneer of alluvium, as in other basins in Wyoming. That the gravel is somewhat thicker in Jackson Hole than elsewhere may be due to the fact that glacial waters from the Jackson Lake moraine poured a large volume of outwash southward along Snake River. (See Fig. 27).

The facts which directly suggest that the fault is an old one, and the scarp due merely to the great contrast in the resistance offered by the rock masses on the east and west sides to erosion, are: (*a*) all other faults within 25 or more miles of this region are old faults, which have lost all original topographic expression; (*b*) it is particularly

uplift have been eroded much more rapidly than they otherwise would have been. The Archean exposures of the Owl Creek Range are not large, and probably did not appear until the last two or three thousand feet of denudation were accomplished. We are reduced, therefore, to a consideration of the Wind River Range as the only part of the district in which the Archean rocks are

Fig. 28.—A typical canyon in the hard Carboniferous rocks of the Wind River Range, twelve miles west of Lander.

broadly exposed and under simple conditions. The Bighorn, Sherman, and certain other ranges in the state should probably be put in the same category.

significant that this is true of the Phillips Canyon fault which appears to join the main Teton fault at an acute angle as if it were merely a branch of it; (*c*) sedimentary rocks to a thickness of perhaps more than 10,000 feet have been removed from the Teton, block since it was first elevated, but these are still preserved in the bottom of Jackson Hole; (*d*) the abruptness and height of the escarpment decrease immediately and markedly toward the north and south ends of the range, as quickly as the outcrops of the less resistant Paleozoic and Mesozoic strata are reached; (*e*) the flat-bottomed part of Jackson Hole coincides almost exactly with the distribution of the weakest Eocene clays and the valley becomes narrow again southward as soon as the Cretaceous outcrops are reached.

In the Wind River uplift deep canyons have been carved out of the granitic rocks, but extensive remnants of the old plateau surface are still preserved upon the interstream divides (Fig. 12). The relative resistance of the hard Archean outcrops is well shown in the channels of many of the creeks that flow out from the Wind River Range. Thus it is generally easy to ascend these valleys as far as they have been cut through the sedimentary strata, but from the base of the Cambrian inward to the axis of the anticline,

FIG. 29.—A canyon in the Archean rocks south of Mount Moran, in the Teton Range.

the Archean outcrop is marked by a series of rapids, falls, and narrow gorges which make even foot travel difficult if not impracticable. Along old faults, which have brought the Archean into contact with some of the weaker members of the sedimentary column—as for example south of the great bend of Green River—the streams have cut merely a few notches in the pre-Cambrian rocks during the time in which the softer sedimentary beds have been completely stripped away to a much lower elevation. This satisfactorily explains why the Wind River plateau has an abrupt front in this locality but not elsewhere.

In short, it appears that the Quaternary has been characterized by widespread changes of level, which in turn have induced the erosion of deep valleys, and that these valleys have been widened in proportion to the weakness of the strata in which they were carved (Figs. 28 and 29). Although there is an advantage in stating the history in these simple terms in order to display the general conditions, it is appropriate to say that the events of the Quaternary have been much more varied and complex than this generalized statement implies. There is good evidence that the uplift or other interruption did not take place once for all at the beginning of the period, nor uniformly throughout the period, but at intervals scattered through a long time. Denudation has been accented by these disturbances, and doubtless also by climatic changes, of which mention has not yet been made. Furthermore, the work of denudation has been done, not only by streams, but in minor degree also by winds, glaciers, avalanches, ground-water, and unaided gravity. This complex chapter of the history may now be analyzed; and, since the changes produced by the different agencies are not in all cases readily correlated, I find it best to treat them separately and in the order of their relative importance.

THE MODE OF ORIGIN OF COAL[1]

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Harvard University

The question of the manner of formation of coal has divided geologists for over a hundred years and the problem is still disputed, although at the present moment the weight of geological opinion inclines in one direction. Where so much difference of opinion has existed for so long a period, new evidence is obviously much to be desired. The current views as to the origin of coal are based mainly on a consideration of the stratigraphic evidence other than that supplied by the coal alone. For example, much importance has been attached to the presence of underclays beneath the coal beds, as evidence of the existence of an ancient forest soil. Likewise a considerable amount of importance has been attached to the existence of Stigmarian and other root or rootstock systems in the substratum beneath the coal. Comparatively little evidence has been derived from the consideration of the coal itself. Where such evidence has been available it has been of a contradictory nature. For example, certain coals rich in bituminous matter, such as cannels and the like, obviously present in their organization strong resemblances to the muck formed in the bottoms of modern lakes; that is, they consist, to a very large extent, of the remains of spores and pollen. In other cases, much more rare, concretions have been found in the coal, containing quantities of often well-preserved vegetable remains. These remains are of such a character and are related to one another in such a way, that their clearest modern representative is the peat of existing peat bogs. The two kinds of evidence derived from petrified coals on the one hand or from the persistent structures in unmineralized coals on the other appear to justify diametrically opposite views as to the conditions of coal formation. Similar contradictions

[1] *Contributions from the Phanerogamic Laboratories of Harvard University, No. 67.*

apparently present themselves, when the strata below (underclays, etc.) or above (roof strata) the actual coal beds are considered. We have as a consequence the two opposed schools of opinion in regard to the origin of coal. The hypothesis which homologizes the coal bed with an existing peat deposit is known as the autocthonous or *in situ* theory. On the other hand, the view that coal represents an accumulation by floating and sedimentation in the bottom of open lakes or lagoons is known as the transport or allocthonous hypothesis. The evidence for and against these two opposed views has recently been so admirably summarized by Professor John J. Stevenson in his *Formation of Coal Beds* that it appears unnecessary to refer to the subject further at this time.

The purpose of the present communication is to emphasize the importance of the study of the coal itself, in connection with any views as to its composition and mode of origin. The investigation of coal by means of the microscope has presented many difficulties, chiefly because of the opacity and texture of the coal itself. The methods which have been successful in the case of other minerals have not given satisfactory results in the case of coal. The thin ground sections of the mineralogist and petrologist are of little value when the material investigated is coal. If the ground sections are made thin enough to show all of the structural features, these are destroyed by the abrasive effect of the process on the comparatively soft and friable substance of the coal itself. The writer has had considerable experience in investigating the remains of fossil plants in a carbonized condition, a state of preservation which has hitherto been neglected in favor of petrified material. Material in a mineralized condition of preservation is, however, comparatively rarely available, while carbonized vegetable remains are often abundant. A modification of the writer's methods in the case of carbonized Mesozoic plants has been developed as the result of a long series of experiments, and it is now possible to secure satisfactory thin and transparent sections on the microtome of practically all kinds of coals. Obviously this possibility put the question of coal composition and mode of formation on an entirely new footing, as it is now possible to investigate these subjects satisfactorily from the standpoint of the organization of coals of various

geological ages in different parts of the world. The writer has examined coals derived from regions as far apart as Alaska and Patagonia, Dakota and Texas, Washington and Virginia, Great Britain and Japan, Sweden and Tasmania in geographical separation, and in stratigraphic distribution from the Devonian to the Miocene and the present epoch. The results of these investigations have been satisfactorily uniform. We are now apparently in the position to judge of the great problems of coal formation from the all-important aspect of the organization of the coal

FIG. 1

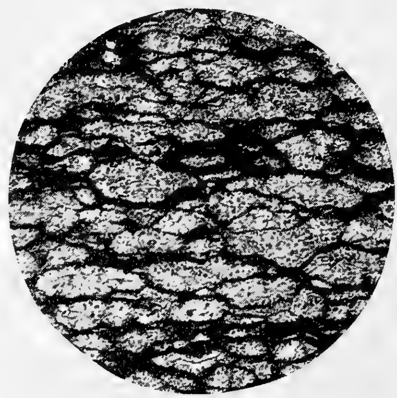

FIG. 2

itself. In other words, we are no longer compelled to infer the nature of coal from the conditions of deposit of the surrounding strata, which may have been laid down under very different circumstances, but we can reach definite conclusions from the structures revealed by the coal itself under the microscope.

Reference has already been made to the fact that certain coals, containing large quantities of spores, are generally admitted to have been formed, not on land as the case with peat, but as sedimentary deposits in open water. Fig. 1 illustrates a notable carbonaceous deposit of this type, namely tasmanite. In the upper part of the figure one spore still retains its rounded contour, but the spore material as a whole, which is very abundant, is represented by the collapsed spore coats, inclosing a linear cavity. Fig. 2 shows the structure of another type of spore coal, an oilshale from

Australia. Here all the spores have collapsed and have rough alveolar exteriors. A remarkable feature of oilshales is the very large amount of spore material present in them, a condition correlated directly with their rich petroleum content when subjected to distillation. A curious error of interpretation has been made in regard to oilshales, bogheads, or bituminous schists, as they have been variously termed. Their structural elements have been regarded by Renault, Bertrand, and Potonié as the remains of delicate gelatinous algae. The present writer, by the improved methods already referred to, has secured evidence that the structures in question are the very rough spores of extinct fernlike cryptogams.[1] It is clear, further, that they cannot be anything of so delicate a nature as algae, since highly modified wood is frequently found in the same matrix with them. It is obvious that gelatinous algae would have quickly been obliterated by influences so powerful as to eliminate the structural organization of wood.

FIG. 3

Fig. 3 shows the organization of Kentucky cannel. Scattered throughout the matrix are light bodies which are yellow in the coal itself. These represent spores. They are inclosed in a darker matrix which in some cases reveals lighter, somewhat undulating or crinkled bands or particles. These structures are brown in color in the coal and represent the remains of wood. A large stripe of this kind runs nearly across the illustration. Cannel owes its highly bituminous character to the presence of large quantities of spores, just as is the case with oilshales and tasmanite. Since the spore material is less pure than in the former types of coals, the bituminous matter is less and the distillate of petroleum is correspondingly reduced in amount.

[1] "The Nature of Some Supposed Algal Coals," *Am. Acad. Arts and Sci.*, XLVI (December, 1910).

It is universally admitted that all three types of coal shown in the illustrations are formed as a result of water deposit. It will be convenient at this stage to consider rather rare types of coal which show considerable evidence of being derived from ancient peat beds. Coals of this organization represent old swamps skirting the sea and correspond, with certain necessary allowances made for the difference of vegetation in the Paleozoic, to our existing tidal mangrove estuaries. They are usually designated on this account as paralic (*paralisch* of Naumann) coals. In some coals of this type from Great Britain, Westphalia, and the Donetz coal field of Russia

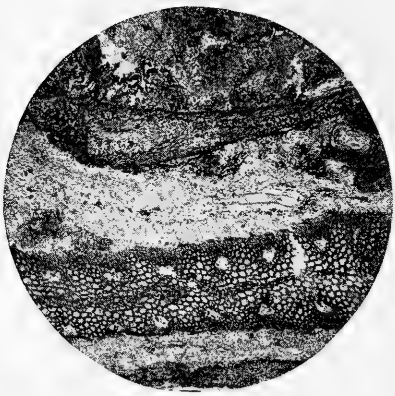

Fig. 4

Fig. 5

concretions have been found known as coal balls. A very good account of these structures, particularly for the English examples, was published, not long since, in the *Philosophical Transactions of the Royal Society of London.*[1] Our Fig. 4 shows the structure of a coal ball from the Westphalian coal field in Germany, which I owe to the courtesy of the Geologische Landesanstalt of Prussia. In the lower part of the figure is the crushed stem of a young calamite and the rest of the area is made up of less recognizable remains. Fig. 5 illustrates the appearance of the coal adhering to the outside of the same coal ball. It has been shown clearly by

[1] Stopes and Watson, "On the Present Distribution and Origin of the Calcareous Concretions in Coal Seams, Known as 'Coal Balls,'" *Phil. Trans. Roy. Soc. London*, Series B, Vol. CC, 167–218.

Stopes and Watson that true coal balls are formed *in situ* in the coal by local impregnation of the original peaty substance with lime or magnesium carbonate or both. The mineralized portion of the peat retains its structure, while that which has not been impregnated is completely devoid of definite organization when transformed into coal. If all coals were derived from ordinary peat they would present the same homogeneous organization as shown in our Fig. 5, since wood and similar formations have their structure obliterated when transformed into coal. In this respect peaty deposits present a marked contrast to the lacustrine mucks which have been the forerunners of cannels and oilshales. The spores and similar cutinized structures are the only remains of plants, which in general do not become obliterated in organization when passing into coal. Only petrification can save ordinary peat in the long run from the complete obliteration of its original vegetable constituents.

Fig. 6

It is now possible to consider with advantage the commoner types of coal, those which occur in greatest abundance in all parts of the world where coal deposits exist. Fig. 6 illustrates a coal of this ordinary bituminous type from Illinois. It is at once clear that the structure is not homogeneous as in the Westphalian coal shown in Fig. 5. Running across the figure is a dark band of "mother of coal" or mineral charcoal. This represents wood which was charred by fire before it entered into the substance of the coal. Mother of coal, inappropriately so called, represents the only woody constituent of ordinary coals which retains its structure. Very often the woody elements are in a perfect condition of preservation and the nature of the trees which produced them can be diagnosed. In addition to the dark band of mineral charcoal or mother of coal, there are other narrower dark stripes in the coal, as well as lighter-hued more homogeneous bands. The latter correspond to highly

modified wood. The nature of the others can best be understood from Fig. 7. This represents a portion of the last figure, much more highly magnified. The dark stripes are now seen to contain numbers of very light-colored bodies or spores.

Fig. 8 shows the organization of a very highly bituminous coal from Kentucky. The crenulated bands corresponding to layers of modified lignitic substance stand out very clearly in this illustration. Between the woody or lignitoid zones are darker stripes variegated with light linear bodies. The latter are compressed

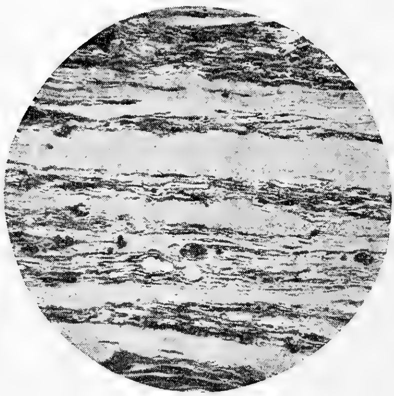

Fig. 7

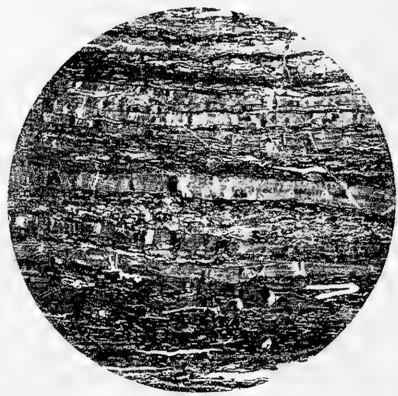

Fig. 8

spores. Fig. 9 shows a part of the foregoing highly magnified, and the canneloid nature of the darker bands alternating with the lignitoid layers can now be clearly ascertained.

Further examples of coals may now be considered. Fig. 10 shows the structure of a bituminous coal from European Russia. Clearly it is largely composed of spores. Through the courtesy of M. Zalessky of the Comité Géologique of St. Petersburg, I have had the opportunity of examining a considerable number of coals from European as well as Asiatic Russia, with similar results as regards organization (except of course the coals of paralic origin). Fig. 11 reveals the structure of non-coking bituminous coal from Indiana. The conditions presented are the same as in the case of the Illinois coal, with the exception that the spore material is very scanty, as indeed is the case with all lean bituminous coals, in

contrast to fat coking and gas coals. Fig. 12 illustrates the organization of a bituminous coal from St. Caterina, South America. The spores and lignitoid substances are as characteristically present as in the other cases. Fig. 13 shows the structure of a high-grade steam coal from Virginia. Here, as in the other cases described for bituminous coals, spores as well as lignitoid substance are found. As a final illustration, in Fig. 14 is shown a coal from Lancashire in England. The narrow stripes of lignitoid and the abundant light-hued spores can readily be discerned.

Fig. 9

Fig. 10

Through the courtesy of the Director of the United States Geological Survey and Dr. David White, Chief Geologist, I have had the opportunity of examining nearly two hundred coals from various states of the Union and from Alaska. The results of investigation so far as bituminous coals are concerned, from whatever geological horizon obtained, showed in all instances the presence of a greater or less amount of spore material. Through the kindness of the Director of the Geological Survey of Canada, I have been able to examine into the structure of the Mesozoic coals of the Canadian West and Vancouver Island, as well as the coking and non-coking Carboniferous and Permian coals of New Brunswick, Nova Scotia, and Cape Breton Island. The results here were similar to those obtained in American coals. The coal deposits of England and Scotland have also supplied numerous and varied

samples, which in all cases led to the same conclusion: namely, that spores are a practically unfailing constituent of ordinary coal. The coals of Scania (Triassic) in Sweden have yielded results in harmony with those already described for other countries and continents. Of oriental coals I have, through the courtesy of the United States Bureau of Mines, been able to examine several from Japan and China. These revealed the same unfailing spore content characteristic of other material examined. I have also examined a number of coals from both the Queensland and New South Wales

FIG. 11

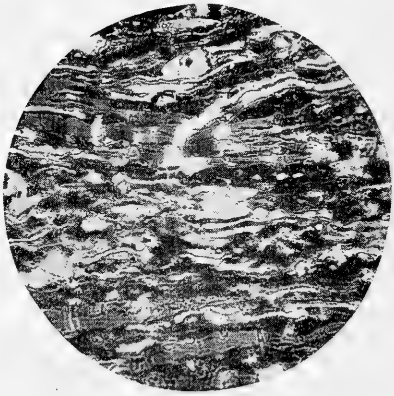

FIG. 12

coal fields of Australia, which in some cases showed a very high spore content and in all instances contained these bodies. Coals from Venezuela, Chile, Brazil, and Patagonia of various geological periods yielded similar results on microscopic investigation.

The almost universal presence of quantities of spores in coals is a phenomenon not without significance in connection with our views in regard to the origin of coal substance. In the case of cannels, oilshales, and similar coals, the high spore content is interpreted on all hands as clear evidence of their formation in open water. In contrast, however, to the views held in regard to the origin of cannels and coals of similar organization are those entertained in regard to the conditions of formation of ordinary bituminous coals. These coals are regarded by the majority of geologists and paleobotanists, both in Europe and in this country,

as of terrestrial origin. The reasons for this belief are drawn mainly from the strata related to the coal beds, but it is somewhat generally admitted that the character of the underclay or seatearth of a coal bed is no more final proof of its condition of origin than the cover of a book is necessarily convincing evidence as to the age and nature of its contents. It becomes clear, as a result of the microscopic examination of coals at large, that they cannot in general be derived from ordinary peat, as they do not show the structure of coals which by inclosed petrifications are known to be of peaty deriva-

FIG. 13

FIG. 14

tion. The paralic or old seaboard coals of Great Britain and parts of Continental Europe have a homogeneous structure, as has been shown above in the case of Westphalian coals. This homogeneity is not at all due to opacity of the thin sections, because even when quite translucent, as when prepared by the improved methods devised for the sectioning of coal, preliminarily described in the article on "Supposed Algal Coals," cited earlier in the present communication, they show no clear evidence of structure, except so far as they contain fragments of carbonized wood, a condition not uncommon in the paralic coals of British origin.

Since it is generally admitted that coals containing large quantities of spores are of aquatic origin, that is, are deposited in open water, the question arises naturally, in view of the conditions described above for ordinary bituminous coals, whether they are

not also laid down in open lakes, lagoons, or tranquil estuaries. There seems to be little doubt that this is the case, because not only do they show to a large extent the organization of cannels but they often pass imperceptibly into coals of this type. Moreover, an equally cogent argument is supplied by the examination of lacustrine muck of the present age. In the greatest depths of our modern lakes, which are usually of glacial origin, one ordinarily finds quantities of diatomaceous tests, or, if the region be a limestone one, a thicker or thinner stratum of a white limy deposit, containing the remains of stoneworts or Characeae. Above the basal deposits occur strata extremely rich in the pollen of conifers and certain catkin-bearing plants, such as the willow, the alder, and the birch. As the water grows shallower by accumulation of organic mud, we begin to find remains of water-lilies and other aquatics mingled with pollen. Then come layers of coarser remains of land plants, twigs, leaves, bits of wood, etc., swept by rain and winds into the lake and always more or less intermingled with a decreasing proportion of spores, until the organic silting of the waters results in the emergence of land, which if not characterizable as dry, still is able to support the growth of land plants. With the use of a peat prober of the type devised by Dr. C. A. Davis, of the United States Bureau of Mines, it is possible to recognize, by probing the depths of the bogs which vegetate on the bosoms of filled lakes, the general order of deposits outlined in the foregoing sentences. Ordinarily the depth of one of these obliterated lakes is not more than ten to twenty feet, so that we have the whole interval from the ice age to the present for the accumulation of the filling material. The estimates of the remoteness of the last glaciation vary from 25,000 years to four times that length of time. Even accepting the smallest estimate, the interval required to fill a lake ten feet in depth is great enough to call for a deposit of less than one foot of material in a thousand years, or not more than one-tenth of a foot in a century. This extremely slow rate of deposition relieves us from having to answer a stock objection to the possibility of the lacustrine formation of the raw materials of coal. It has often been pointed out that a rate of erosion necessary to fill up a large body of water in a relatively brief interval,

with organic materials, would be torrential in its action. As a matter of fact, judging from actual conditions, given abundance of time, almost unlimited accumulation of vegetable remains in water, without any appreciable admixture of mineral matter, can be accounted for, particularly since vegetable substance once permanently covered by water is practically imperishable.

A very serious objection to the peat hypothesis of the origin of coal lies in the observed fact that peat bogs are not found in the parts of our present world which in their climatic warmth correspond to the probable conditions obtaining during the great coal age, the Carboniferous. Vegetable matter not continuously covered by water quickly disappears in tropical climates. True, Potonié has called attention, in his truly monumental work[1] on peat and similar formations, to the existence of what he considers to be a true peat bog in central Sumatra. The illustrations given show clearly, however, that there is present a filled body of water with trees growing upon its scarcely elevated surface. The organic material, moreover, as figured in Potonié's work, contains large quantities of pollen, thus showing its essentially open-water origin. There can be no reasonable doubt that it was as impossible for vegetable matter to accumulate on land, even on wet land, in the Carboniferous and later periods in the Mesozoic and Tertiary, as it is in subtropical and tropical climates at the present time.

We are then apparently brought to the conclusion, both from the conditions of peat formation in the present age on the one hand, and from the actual organization of the commonest and most abundant coals of every geological age and in all parts of our earth, on the other, that the raw materials of coal were heaped up, not as the result of the growth of successive generations of plants on the prostrate and persistent bodies of those already fallen, but as the age-long gradual accumulation of vegetable matter in open water. In other words, coal is not a compost heap but a sedimentary deposit. Striking as are our actual peat bogs in temperate climates, in reality they throw no light on the accumulations of vegetable matter in past ages, which have been preserved to us in the form of coal. The less conspicuous accumulations of vegetable matter

[1] *Die Recenten Kaustobiolithe u. ihre Lagerstaetten*, Berlin, 1908–12.

in the bottoms of our existing lakes afford the real elucidation of the mode of formation of coal beds. As in our actual bodies of water, which accumulate vegetable deposits, the nature of the material laid down varies with the depth, tranquillity of the water, and other conditions, so the nature of coals derived from similar formations of past ages varies in composition. Where the spore material is more abundant, the result is a cannel or an oilshale. Where it is still plentiful but not superabundant, we have coking and gas coals as a product. Where poverty is manifest in the spore content, lean or ordinary bituminous coals are formed. The processes of devolatilization leading to the formation of anthracites and similar high-grade coals will not be considered in the present connection.

CONTRIBUTIONS TO THE PETROGRAPHY OF JAVA AND CELEBES

JOSEPH P. IDDINGS AND EDWARD W. MORLEY

In a short visit to Java and Celebes in 1910 one of the authors visited several localities in Java, the rocks of which have been described by Verbeek,[1] and also the region of the Pic de Maros in Celebes, from which rocks were collected by Paul F. Sarasin and subsequently were described by C. Schmidt.[2] Part of the material collected on this visit has been analyzed chemically by the other author of this paper, and the analyses have been published, for the most part, without special description in the second volume of *Igneous Rocks* recently printed.[3] It is the purpose of this paper to call attention more specifically to the characters of the rocks analyzed, and to point out certain chemical and mineral relationships between the leucitic lavas of Mt. Mouriah in Java and the shonkinites and nephelite syenites in the vicinity of the Pic de Maros in Celebes.

Mt. Mouriah in the Diapara Residency, northeast of Semarang, on the north coast of Java has been described by Verbeek as an extinct volcano, 1595 m. high, which has undergone extensive erosion but still exhibits evidences of two large circular craters. Associated with it are two smaller volcanoes of similar character, Paliaïan and Tülering. The rocks of these mountains are exclusively leucitic, but some varieties are poor in leucite. Other leucitic lavas occur in the small volcano Lourous, and the larger, much eroded volcano Ringgit, in Besouki Residency in Eastern Java. Leucitic rock, together with phonolites, also occurs on the island

[1] R. D. M. Verbeek and R. Fennema, *Java et Madoura*, Amsterdam, 1896.

[2] P. and F. Sarasin, *Insel Celebes; Anhang, Untersuchung einiger Gesteinssuiten.* Wiesbaden: C. Schmidt, 1901.

[3] J. P. Iddings, *Igneous Rocks*, II. New York, 1913.

Bawéan, between Java and Borneo. Verbeek states that Mt. Mouriah is in large part tuffs and breccia with some flows of massive lava. He describes the rocks as chiefly leucite tephrites and leucitites with small amounts of feldspar in the groundmass; there being all gradations between these varieties. Olivine-bearing varieties are much fewer. In some of the rocks there are phenocrysts of leucite 5–10 mm. in diameter, and in exceptional instances 15 mm. The groundmass of these rocks is in most instances dark gray to black, dense and aphanitic, less often porous.

Two of the localities on Mt. Mouriah mentioned in Verbeek's description were visited in 1910: one was the stream Kali Gillinan, near the village Masin, on the south slope of the mountain above Bareng; the other locality was the stream Kali Sekatak, below the village Ragou, above Petjangaän, at the west base of the mountain. The first stream has a narrow channel at the place visited, and washes great bowlders of lava, and has short beaches of smaller bowlders and gravel. The second stream is much larger, flows in more open country below Ragou, and has long reaches of gravel and bowlders. In both places the bowlders represented great varieties of leucitic rocks which were mostly dense and compact and extremely fresh and unaltered by weathering, leucite crystals at the surface of the bowlders being transparent, or only slightly whitened in some instances, but completely altered in others. However, when the rocks are broken the interior portions usually appear to be very fresh, which proves to be the case when their sections are studied microscopically. These leucitic rocks, although considered by Verbeek to be possibly of late Tertiary age, are as fresh and as well preserved as modern lavas, or as many Tertiary lavas in the arid regions of Western America. In places where they have been covered by soil and vegetation for long periods of time they are completely decomposed for a short distance from the rock surface, as Mohr has shown to be the case with andesitic rocks in other parts of Java. Of 37 specimens collected from various bowlders at the two localities named, 8 varieties have been analyzed chemically, the analyses and norms being given in Table I, together with analyses of lavas from two active craters in other parts of Java; a short petrographic description of each follows.

TABLE I

Analyses of Lavas from Java

	1	2	3	4	5	6	7	8	9	10
SiO_2	51.85	50.18	48.32	48.66	47.73	46.54	46.60	45.03	55.42	51.12
Al_2O_3......	19.08	17.82	17.81	17.69	17.93	15.95	16.73	16.59	17.39	19.59
Fe_2O_3......	4.25	4.04	4.65	4.66	4.47	5.24	4.17	4.55	1.56	2.86
FeO......	2.69	3.89	4.62	4.40	4.58	5.51	4.78	6.37	6.82	6.53
MgO......	1.48	2.88	3.37	3.03	4.27	4.70	4.65	3.95	3.28	4.47
CaO......	5.81	7.19	9.15	6.43	9.59	10.09	10.82	11.09	7.57	9.54
Na_2O......	4.46	3.29	3.14	3.93	3.62	2.28	2.62	3.53	2.41	3.11
K_2O......	6.61	6.65	4.79	6.10	4.81	4.44	5.47	5.29	2.67	0.57
H_2O+....	0.55	0.96	0.82	0.80	0.44	0.52	0.71	0.34	0.17	0.11
H_2O-....	0.49	0.55	0.17	0.58	0.24	0.59	0.45	0.15	0.00	0.00
TiO_2......	0.66	0.76	0.88	0.81	0.86	1.11	0.95	1.10	1.07	0.86
ZrO_2......	0.00	N.p.	0.00	0.00	0.00	0.01	N.p.	0.00	0.00	0.00
CO_2......	......	......	......	......	......	......	......	......	......	......
P_2O_5......	1.23	0.76	0.82	0.79	0.52	1.18	1.50	0.96	0.58	0.14
Cl........	0.21	0.16	0.10	0.24	0.17	0.07	0.08	0.26	0.11	0.10
F........	0.10	0.02	0.04	0.16	0.07	0.06	0.17	0.11	0.03	0.04
S........	0.01	0.02	0.23	0.05	0.04	0.09	0.01	0.05	0.03	0.06
Cr_2O_3......	0.00	N.p.	0.00	0.00	0.00	N.p.	N.p.	0.00	0.00	0.00
MnO......	0.51	0.30	0.41	1.49	0.96	0.18	0.41	0.64	0.71	0.65
BaO......	0.17	0.25	0.11	0.16	0.10	0.13	0.21	0.16	0.13	0.03
SrO......	0.19	0.29	0.21	0.21	0.17	0.24	0.13	0.16	0.03	0.03
	100.40	100.01	99.64	100.19	100.57	99.53	100.46	100.33	99.98	99.81

Norms of the Lavas Analyzed

	1	2	3	4	5	6	7	8	9	10
Q........	......	......	......	......	......	......	......	......	7.9	2.2
Or.......	38.9	39.5	28.4	36.1	28.4	26.1	28.4	14.5	16.1	3.3
Ab.......	18.3	7.9	11.0	8.9	3.7	5.8	......	......	19.9	25.7
An.......	13.3	14.2	20.3	13.3	18.9	20.3	17.5	15.3	28.9	38.1
Ne.......	9.7	10.8	8.5	12.5	13.9	7.4	11.9	14.8	......	......
Lc.......	......	......	......	......	......	......	3.5	13.1	......	......
Hl.......	0.4	......	......	0.4	0.2	......	......	0.5	0.1	0.1
Di.......	7.1	14.1	16.6	11.1	20.8	23.1	21.3	27.9	4.6	6.9
Hy.......	......	......	......	......	......	......	......	......	16.4	17.1
Ol........	1.1	2.5	2.9	5.8	4.2	3.7	4.3	2.6	......	......
Mt.......	6.3	5.8	6.7	6.7	6.5	7.7	6.0	6.7	2.3	4.2
Il........	1.4	1.5	1.7	1.5	1.7	2.1	1.8	2.1	2.1	1.7
Pr	......	......	0.4	......	......	......	......	......	......	......
Ap.......	2.7	2.0	2.0	2.0	1.3	2.7	3.7	2.4	1.3	0.3

1. Vicoite, borolanose-monzonose, 'II. 5(6). 2. 3. Gillinan, near Masin. E. W. Morley.
2. Leucite tephrite, borolanose, II. '6. 2. '3. Near Ragou. E. W. Morley.
3. Vicoite, —shoshonose, II. 5(6). '3. 3. Near Ragou. E. W. Morley.
4. Orthoclase-bearing leucite tephrite, borolanose, II. '6. 2. 3. Gillinan. Near Masin. E. W. Morley.
5. Biotite vicoite, borolanose—, II, '6. (2)3. 3. Near Ragou. E. W. Morley.
6. Leucitophyre, kentallenose, 'III, 5. '3. 3. Gillinan, near Masin. E. W. Morley.
7. Leucite tephrite, ourose, (II) III. 6. '3. '3. Near Ragou. E. W. Morley.
8. Vicoite, cascadose, 'III. 7. 2'. 3. Near Ragou. E. W. Morley.
9. Glassy shoshonite, shoshonose, II. (4) 5. 3 (4). 3'. Bromo Crater. E. W. Morley.
10. Basalt, hessose, II. 5. 4. '5. Goentoer lava. E. W. Morley.

The rock whose analysis is No. 1 is a light-gray, aphanitic variety from Kali Gillinan near Masin, which contains rather numerous large phenocrysts of leucite, ranging in size from a diameter of 10 mm. to microscopic dimensions. In thin section they exhibit characteristic polysynthetic twinning, and the index of refraction has been determined by Dr. F. E. Wright to be 1.507 for sodium light. Their chemical composition is shown by the following analysis by Morley:

SiO_2	54.97
Al_2O_3	22.21
K_2O	19.98
Na_2O	0.81
Fe_2O_3	0.61
MgO	0.26
CaO	0.49
H_2O-	0.08
H_2O+	0.56
	99.97

The groundmass is minophyric and dopatic with abundant small phenocrysts of leucite, augite, and some minute feldspars. In thin section it is seen to consist of small phenocrysts of euhedral leucite, zonal augite, and numerous zonal crystals of calcic andesine or labradorite in a holocrystalline groundmass composed of leucite, prismoids of greenish augite, anhedrons of magnetite, with prismoid plagioclase and anhedral orthoclase, some of which form narrow shells around the plagioclase, as in some shoshonites. The calculated norm shows 38.9 orthoclase, 18.3 albite, and 9.7 nephelite, whereas the mode, or actual mineral composition of the rock, shows that no nephelite crystallized from the magma, but considerable albite, which appears in the lime-soda feldspar. Readjusted on this basis the rock might contain about 1.1 orthoclase, 36.2 albite, 29.7 leucite, and no nephelite. Owing to the smallness of the crystals in the groundmass an actual measurement of the mode of the rock cannot be made without great difficulty. The rock may be called a vicoite, which is equivalent to a leucite shoshonite.

The rock of analysis No. 2 is medium gray, very similar to that of No. 1, and occurs on Kali Sekatak near Ragou. The phenocrysts of leucite range in size from 10 mm. downward and in places are

partly separated from the matrix by small spaces or cracks, which indicate that the magma was so stiff just before it solidified that it pulled apart through stretching. In thin section there are small phenocrysts of leucite, augite, magnetite, and reddish-brown biotite, which has an outer zone with weaker absorption than the central part, and in some instances has a margin filled with inclusions. The groundmass consists of leucite, slender prisms of lime-soda feldspar, minute prisms and anhedrons of augite, and magnetite. The chemical analysis and norm show that the rock is a little more femic than the variety first described, No. 1, and that it contains slightly less soda and normative albite. In place of 39.5 per cent of normative orthoclase and 10.8 of normative nephelite, there has crystallized abundant leucite and no nephelite. The rock is leucite tephrite.

The rock of analysis No. 3 is gray, dense, vitreous to subvitreous, and aphanitic, and is from Kali Sekatak. It is minophysic and dopatic, with abundant small phenocrysts of augite. In thin section the groundmass is seen to be holocrystalline and to consist of lime-soda feldspar surrounded by alkalic feldspar, besides anhedral crystals of alkalic feldspar and some leucite, with minute crystals of augite and magnetite. The chemical analysis and norm show this to be a still more femic variety with less alkalies. No nephelite or olivine is recognizable in the rock, and none was observed in Nos. 1 and 2. The rock may be classed as a vicoite, or an orthoclase-bearing leucite tephrite.

The rock of analysis No. 4 is light gray, and aphanitic, from Kali Gillinan. It is dopatic, with small phenocrysts of augite, but no megascopic crystals of leucite or feldspar. In thin section it is seen to contain numerous small phenocrysts of augite, leucite, and very small magnetites, in a holocrystalline matrix composed of prismoid lime-soda feldspar, anhedral orthoclase, leucite, and augite. The chemical analysis and norm are much like those of No. 2, which, however, is characterized by large phenocrysts of leucite. The rock is a variety of leucite tephrite with orthoclase, related to vicoite.

The rock of analysis No. 5 is medium gray, aphanitic, and very porous, from Kali Sekatak, and has some large phenocrysts of

biotite, ranging downward to microscopic crystals, also phenocrysts of augite, but none of leucite or feldspar. In thin section there are abundant small phenocrysts of zonal pleochroic augite, considerable brown mica, zonal lime-soda feldspar, and magnetite, with relatively large crystals of colorless apatite, and a few anhedral, brownish-green hornblendes. There is some ill-defined colorless mineral which may be in part orthoclase, in part leucite. No olivine is to be seen. The chemical composition is very similar to that of No. 3. The norm shows it to be slightly more femic, with more normative nephelite and olivine, and the same amount of normative orthoclase, but the crystallization of abundant biotite accounts for the non-appearance of olivine, and the small amount of leucite without nephelite in the mode. The rock might be classed as a variety of biotite vicoite.

The rock of analysis No. 6 is dark gray and dense, from Kali Gillinan. It is minophyric, sempatic, with comparatively few large phenocrysts of leucite, and abundant small phenocrysts of augite and leucite. In thin section there are abundant leucites, as phenocrysts, as clusters of crystals, and as microscopic crystals in the groundmass. Some clusters of leucites surround augite, others inclose plagioclase, augite, and magnetite. The phenocrysts of augite have pronounced zonal structure; there is a small amount of euhedral olivine. Small crystals of magnetite in some instances have very irregular outlines owing to pockets of groundmass and partly inclosed minute crystals of augite and plagioclase; the outline of the magnetite being rounded in places, as is the case with quartzes having similar partial inclusions. The rounded forms are clearly forms of growth and not of solution, as sometimes suggested. The holocrystalline groundmass contains considerable zonal calcic plagioclase, besides minute crystals of the other mineral constituents of the rock. The chemical analysis and norm show that this variety is more femic than the preceding ones, and that it is richer in normative anorthite. Normative orthoclase and nephelite are represented by modal leucite and by albite in the lime-soda feldspar. Olivine is modal as well as normative. The rock may be called a leucitophyre or leucite basanite.

The rock of analysis No. 7 is gray, dense, and aphanitic with numerous pores, and is found in Kali Sekatak. It has abundant

small phenocrysts of augite, which in thin section are seen to be zonal. The groundmass consists of abundant small leucites with an interstitial matrix composed of augite, magnetite, and minute prismoids of plagioclase. The chemical analysis and norm are very similar to those of No. 5, but the mode differs in having much leucite and no biotite. The rock is a leucite tephrite.

The rock of analysis No. 8 is gray, dense, aphanitic, and non-porphyritic, from Kali Sekatak. In thin section it is seen to be holocrystalline, and to consist of prisms of augite with some magnetite, and many leucites with anhedral orthoclase and anhedral alkalic plagioclase, or lime-soda feldspar surrounded by orthoclase. From the chemical analysis and norm it is seen to be the most femic variety from this region that has been analyzed. There are 13 per cent of normative leucite, and nearly 15 of normative nephelite which does not appear as modal nephelite, but must be represented by albite molecules, and by a greater amount of leucite than appears in the norm. The rock may be considered a variety of non-porphyritic vicoite.

While no crystals of nephelite have been recognized in any of the thin sections of these rocks, some of the specimens collected have numerous cavities with small white hexagonal crystals with basal planes, which appear to be altered nephelite. In other specimens there are cavities containing brilliant square prismatic crystals terminated by pyramidal planes over the edges of the prism, which have the index of refraction and habit of stilbite.

In a region where there are such highly potassic lavas as those of Mt. Mouriah it is to be expected that the lavas of more recent date should contain notable amounts of potash. Through the kindness of Dr. Verbeek a study was made of the thin sections of rocks collected by him and deposited in the Bureau of Mines, in Batavia, in order to learn whether orthoclase-bearing varieties of the andesitic lavas could be found. The collection contains few holocrystalline rocks which show orthoclase borders around prismoids of plagioclase, which might be called shoshonites or trachy-andesites. In most cases studied the rocks have a glassy matrix which might contain whatever orthoclase molecules were present in excess of those entering plagioclase crystals. However, the strongly porphyritic glassy lava which occurs in ejected blocks

at the cinder cone of the Bromo volcano in the old Tengger crater, in Eastern Java, has been analyzed chemically with the result shown in analysis No. 9, and proves to be a glassy variety of shoshonite, which shows no orthoclase, or other potassic mineral, in thin section. The dull glassy matrix contains abundant large phenocrysts of calcic plagioclase, having an index of refraction, β, which is 1.560, corresponding to labradorite, $Ab_1An_{1.2}$, and much fewer of brown vitreous augite. In thin section the groundmass is seen to consist of brown glass full of prismoid plagioclase, equant anhedral augite, and magnetite, with phenocrysts of labradorite and a few of augite, olivine, and magnetite.

The recent basaltic lava of Goentoer volcano has been analyzed also, and found to be hessose, with low potash, analysis No. 10. The lava is dark gray, aphanitic, and porous; is minophysic with few large phenocrysts of glassy feldspar with $\beta=1.575$, which are anorthite, Ab_1An_{10}, and many small ones of less calcic feldspar and glassy yellow olivine. The groundmass contains a small amount of globulitic glass base, between abundant microlites of labradorite and fewer of augite, olivine, and magnetite.

The Pic de Maros is a mountain of igneous rocks covered with vegetation, which forms the southwestern extremity of a short ridge, situated between Maros and Tjamba, north of Makassar, in Celebes. Its rocks are exposed in place in a few localities, but may be seen in great variety in bowlders in the stream Gentungen, in the vicinity of Beleangin, and in loose material in drainage channels at the north base of the mountain. From the last two localities numerous specimens were collected by P. and F. Sarasin and afterward were described by C. Schmidt. In a hurried visit made by one of the authors of this paper some additional observations were made of rocks in place along the road on the west and south flank of the mountain, and some other varieties of rock were collected from the stream Gentungen.

About 5 miles up the road from the rest house, Patinoean, toward Tjamba there are large rounded exposures of massive shonkinite, which is exposed again a mile farther on the same road, and at other places. The rocks are dark colored, medium grained, and are almost perfectly fresh on the rough weathered surface. The

freshness of these rocks along the roadside where not covered by soil and vegetation is extremely interesting and was unexpected, since the rocks are intrusive bodies which have been uncovered by gradual erosion, and are not recent lava flows. They consist of abundant crystals of black mica, the largest 4 mm. in diameter, with euhedral prisms of augite, the largest being 6 mm. long, besides nearly equal amounts of glassy feldspar, some of which is prismoid, while others are anhedral. In thin sections these rocks are seen to consist of nearly equal amounts of mafic minerals and feldspar, which are augite, much brown biotite, considerable magnetite and apatite; the feldspar is almost wholly orthoclase in prismoid sections, with very little lime-soda feldspar. One section shows a little interstitial quartz.

Similar varieties of shonkinite occur as bowlders in the stream Gentungen. One variety at this locality is very similar in general appearance to those just described, but is richer in mafic minerals. It is medium grained, in thin section it might be called coarse grained, and consists of much augite and brown biotite, which crystallized almost synchronously and inclose much colorless apatite, and also magnetite. There is also a very small amount of greenish hornblende which has crystallized around augite. The subordinate felsic components of the rock are chiefly orthoclase, or microperthite, which is slightly cloudy, besides some very transparent calcic plagioclase, and a small amount of interstitial isotropic mineral which is probably sodalite. The chemical composition of this variety of shonkinite is shown by analysis No. 16, and its place in the Quantitative System of Classification is found to be in division III. 6. 3. 2, ottajanose, more exactly its symbol is III. 6. (2) 3. 2. The norm contains over 10 per cent of lenads and 12 per cent of olivine which do not appear in the mode, owing to the large amount of biotite which crystallized from the magma. This variety of shonkinite has been called marosite.

Still another variety of shonkinite found in bowlders in Gentungen is characterized by large poikilitic micas, 20–30 mm. in diameter, which lie in all possible positions in the rock. This variety grades into one in which the micas are not poikilitic, but yield brilliant cleavage plates 10–15 mm. in diameter. In thin

section these two varieties of shonkinite show much mica and augite, with considerable apatite and magnetite. The inclusions in the poikilitic mica are augite and plagioclase. There are large poikilitic crystals of orthoclase, and many small crystals of all the constituent minerals of the rock. The chemical composition of this variety is shown in analysis No. 15, which is not very different from No. 16. Owing to somewhat more calcic feldspar in the norm, which also appears in the mode, the rock belongs in kentallenose, III. 5. 3. 3; more exactly it is III. 5 (6). 3 (4). 3, so that it may be called a biotite kentallenite rather than biotite shonkinite.

Another rock in the Gentungen bears a striking resemblance to the pseudoleucite shonkinite, or fergusite, from Montana, described by Pirsson.[1] It has a dark gray, very fine-grained matrix, with abundant equant, whitish pseudoleucites, irregularly scattered and in clusters, the individual spots being about 3 mm. in diameter; a few, 6 mm. There are also some small poikilitic plates of

TABLE II

CHEMICAL ANALYSES OF ROCKS FROM CELEBES, BORNEO, AND SUMATRA

	11	12	13	14	15	16	17	18	19	20
SiO_2	58.79	58.61	56.31	46.08	45.26	43.98	46.04	46.05	61.91	53.75
Al_2O_3	19.55	21.62	21.69	20.40	15.70	12.28	12.40	14.88	16.26	17.06
Fe_2O_3	1.82	1.16	1.20	2.12	2.44	3.49	3.54	4.22	2.45	4.18
FeO	1.43	0.79	0.97	3.27	6.16	7.70	5.58	5.78	3.96	5.50
MgO	0.74	0.16	0.54	6.30	8.28	8.00	12.60	5.98	1.81	4.07
CaO	2.37	1.71	1.88	8.48	11.95	11.19	8.38	13.47	4.35	7.72
Na_2O	4.21	6.60	5.56	2.07	1.73	1.33	1.62	1.41	4.40	3.33
K_2O	8.69	6.82	9.17	6.72	3.42	5.06	4.87	2.56	3.04	1.37
H_2O+	1.05	1.42	1.13	1.70	1.12	1.61	3.55	3.01	0.18	0.50
H_2O-	0.06	0.19	0.00	0.06	0.29	0.12	......	0.52	0.10	0.39
TiO_2	0.54	0.17	0.41	1.39	1.66	2.24	2.20	0.93	0.79	0.88
ZrO_2	0.00	0.01	0.00	0.00	0.01	0.00	......	0.00	0.00	0.00
CO_2	......	......	......	......	......	......	......	......	......	......
P_2O_5	0.11	0.04	0.13	1.19	0.90	1.81	......	0.59	0.40	0.25
Cl	0.12	0.07	0.28	0.10	0.25	0.12	......	0.09	0.13	0.11
F	0.03	0.01	0.03	0.09	0.08	0.15	......	0.03	0.04	0.06
S	0.02	Tr.	0.17	0.06	0.05	0.10	......	0.04	0.05	0.06
Cr_2O_3	0.00	0.00	0.00	0.00	0.00	0.00	......	0.00	0.00	0.00
MnO	0.40	0.40	0.16	0.19	0.34	0.51	Tr.	0.21	0.20	0.50
BaO	0.08	0.01	0.05	0.11	0.10	0.16	......	0.06	0.02	0.02
SrO	0.12	0.02	0.04	0.07	0.06	0.12	......	0.07	0.06	0.12
	100.13	99.81	99.72	100.40	99.80	99.97	100.78	99.90	100.15	99.87

[1] L. V. Pirsson, *U.S. Geol. Survey, Bull. 237*, 1905, p. 74.

NORMS OF THE ROCKS ANALYZED

	11	12	13	14	15	16	17	18	19	20
Q	......	......	......	......	......	......	......	......	13.3	6.8
Or	51.7	40.0	54.5	26.1	20.0	24.5	25.0	15.6	17.8	8.3
Ab	25.7	35.1	14.7	......	3.1	......	......	9.4	36.2	27.3
An	8.6	8.3	8.1	26.7	26.4	12.8	12.2	26.4	16.4	28.1
Ne	5.4	11.1	16.5	9.4	4.8	5.9	7.4	1.4	......	......
Lc	......	......	......	10.5	......	4.4	3.1	......	......	......
Hl	......	......	0.5	......	0.5	......	......	......	0.2	0.1
C	......	0.4	......	......	......	......	......	......	......	......
Di	1.8	......	0.5	6.6	22.0	24.7	23.0	29.5	2.1	7.0
Hy	......	......	......	......	......	......	......	......	7.7	12.9
Ol	1.5	1.0	0.9	10.5	12.5	11.9	17.3	4.9	......	......
Mt	2.6	1.9	1.9	3.0	3.5	5.1	5.1	6.0	3.7	6.0
Hm	......	......	......	......	......	......	......	......	......	......
Il	1.1	0.3	0.8	2.7	3.2	4.3	4.3	1.8	1.5	1.7
Fl	......	......	......	......	......	0.3	......	......	......	......
Ap	0.3	......	0.3	2.7	2.0	4.4	......	1.3	1.0	0.7

11. Trachyte, pulaskose, I'. 5. '2. '3. Gentungen, Pic de Maros. E. W. Morley.
12. Sodalite trachyte, laurvikose-pulaskose, I. 5 (6). (1) 2. 3 (4). Road at S. W. base of Pic de Maros. E. W. Morley.
13. Nephelite syenite, beemerose-procenose, I. '6. (1) 2. 3. Gentungen, Pic de Maros. E. W. Morley.
14. Fergusite, — II. 6. 3. 2'. Gentungen, Pic de Maros. E. W. Morley.
15. Biotite kentallenite, ourose-kentallenose, III. 5 (6), 3 (4). '3. Gentungen, Pic de Maros. E. W. Morley.
16. Marosite, kajanose-ottajanose, III. 6. (2) 3. 2. Gentungen, Pic de Maros. E. W. Morley.
17. Mica-leucite basalt, kajanose, III. 6. 2 (3). 2'. Oeloe Kajan, East Borneo. Pisani and Brouwer.
18. Absarokite, kentallenose—, III. 5.(3) 4. 3. Road at S.W. base of Pic de Maros. E. W. Morley
19. Andesitic pitchstone, dacose, 'II. 4'. 2.' '4. Simaboer, Mesapi, Sumatra. E. W. Morley.
20. Andesite, andose, II. '5. 3'. 4. Rau Rau, Mesapi, Sumatra. E. W. Morley.

brownish-black mica. In thin section the whitish spots are seen to be microscrystalline aggregations of anhedral alkalic feldspar, apparently orthoclase, and anhedral nephelite. The dark-colored matrix resolves itself into a complex of prismoid and anhedral lime-soda feldspar, augite, somewhat poikilitic biotite, magnetite, and olivine. In places around the areas of pseudoleucite there are clusters of minute anhedrons, having the refractive index of pyroxene, which border small crystals of plagioclase or partly replace them. The chemical composition of this rock is given by analysis No. 14, which belongs in II. 6. 3. 2, and is more calcic, so far as normative feldspars are concerned, than fergusose, II. 6. 1. 2. The rock is an equivalent of nephelite monzonite, but has the habit of fergusite.

Besides the shonkinitic rocks in this region there are various syenitic rocks grading into nephelite syenites and their aphanitic

phases, syenite porphyry, bostonite, and trachyte, and also phonolite, which is said to form the summit of the Pic de Maros. There are varieties intermediate between shonkinite and syenite with subordinate amounts of mafic minerals. A nephelite syenite from the Gentungen is light gray and medium grained, with tabular crystals of orthoclase and quite subordinate amounts of mafic minerals. In thin section the feldspar crystals are seen to have a diverse arrangement, and form the bulk of the rock, through which are scattered stout hexagonal prisms of nephelite about the same size as the crystals of aegirite augite, besides smaller amounts of biotite and brown hornblende, also paramorphs of hornblende and mica, and small crystals of sphene. The chemical composition of this rock is given in analysis No. 13, which is in procenose, its symbol being I. ′6. (1) 2. 3.

There are porphyritic trachytes with tabular phenocrysts of glassy orthoclase, some of which are 40 mm. long and 5 mm. thick. A variety with smaller phenocrysts of orthoclase, 15 mm. long, has a bluish-gray aphanitic groundmass and small equant phenocrysts of mafic minerals. In thin section the groundmass is seen to consist of prismoids and anhedrons of orthoclase, with small amounts of sodic plagioclase, nephelite, and interstitial sodalite. The mafic minerals are greenish-brown hornblende having very irregular outline, in part poikilitic with orthoclase feldspar, also small anhedrons of green augite, and magnetite. Its chemical composition is shown by analysis No. 11, pulaskose I′. 5. ′2. ′3. Part of the normative anorthite enters the mafic minerals, and part must be involved in the alkalic feldspar, which must be sodic orthoclase, for no plagioclase feldspar is recognizable in thin section.

On the road from Patinoean to Tjamba, about 7 miles from Patinoean, there is an exposure of altered tuff containing blocks of massive, light-gray, minutely crystalline rock, with few small phenocrysts of glassy tabular feldspar. Under a lens it is seen to consist of subparallel tabular feldspar with abundant reddish spots. In thin section the rock is seen to be holocrystalline, with trachytoid fabric, composed of anhedral crystals of orthoclase, yielding prismoid sections, some of which show Carlsbad twinning.

Scattered through the whole are anhedral, interstitial crystals of isotropic mineral, probably sodalite. These are small amounts of pale-green augite, brown biotite, and magnetite. The rock is sodalite trachyte, or sodalite bostonite, approaching phonolite in composition. Its chemical composition is shown in analysis No. 12 and corresponds to a variety of pulaskose, I. 5 (6). (1) 2. 3 (4). The norm contains 11 per cent of nephelite and 8 of anorthite, but no lime-soda feldspar is recognizable in thin section. Other varieties of aphanitic, more or less porphyritic, trachytes occur on the north slope of the mountain east of the stream Gentungen, some of which contain small amounts of sodalite; others carry considerable biotite.

At the west base of the Pic de Maros there is a sheet of surface lava which is exposed at the Falls of Maros, or Bantinoeran, and also along the road farther south, about 3 miles east of Patinoean, where it is columnar, and is overlaid by limestone. The rock is coarsely porphyritic with abundant phenocrysts of black euhedral augite, the largest 7 mm. in diameter, and less numerous dark-brown glassy olivines, in a dark-gray aphanitic groundmass speckled with small white spots. In thin section the augite phenocrysts are brownish green, slightly pleochroic, and zonal, the central part being lighter colored than the margin. Olivine is abundant and there is much lime-soda feldspar in small phenocrysts, short prismoid and tabular; besides some prismoid orthoclase and zones of orthoclase surrounding plagioclase. There is a fine-grained matrix of the same kinds of minerals with magnetite and zeolitized feldspathoid mineral which may have been analcite, nephelite, or leucite. The rock is a variety of orthoclase basalt, or absarokite, whose chemical composition is shown by analysis No. 18, III. 5. (3) 4.3, a somewhat more calcic rock than kentallenose, which is III. 5. 3. 3.

A comparison of the analyses of rocks from Mt. Mouriah and the Pic de Maros shows that the leucitic rocks of the one and the shonkinitic rocks of the other locality are chemically similar. They are low in silica, high in potash, and relatively high in alumina and calcium oxide. This shows itself mineralogically in the prominence of leucite in the lavas of Mt. Mouriah, and of orthoclase and

biotite in the phanerocrystalline rocks of the Pic de Maros, while augite is a prominent constituent in both series. The striking contrast between the two groups of rocks is the crystallization of leucite in the lavas, and the absence of nephelite, without any considerable amount of orthoclase, which when present is a microscopic constituent of the groundmass, also the absence of biotite in most phases of the leucitic lavas, although in some varieties biotite and orthoclase have been crystallized at the expense of leucite. While in the intrusive rocks of the Pic de Maros there is no leucite, but considerable nephelite in some instances, abundant orthoclase and biotite, and an absence of noticeable amounts of lime-soda feldspar; in some rocks it is probably present moleculary in considerable amounts. These mineralogical contrasts must be due to differences in chemical equilibrium within chemically similar magmas, resulting from physical differences attending the crystallization of lavas in one case and of intruded magmas in the other. Such contrasts have been pointed out before, but never with so good an illustration.

A pseudoleucite kentallenite, or fergusite, occurs with the shonkinitic rocks in Celebes, the former leucites having been replaced by orthoclase and nephelite. Leucitic lavas occur in the neighborhood of the Pic de Maros, and in numerous other localities in southwestern Celebes,[1] and a mica-leucite basalt with very similar chemical composition to the highly mafic shonkinite, marosite, occurs in East-Central Borneo on the Oeloe Kajan.[2] Strongly potassic magmas yielding leucitic lavas, and biotite-orthoclase phanerites, with more or less nephelite, occur widely scattered from Eastern Java, through Southwestern Celebes, and have been found in East-Central Borneo, so that it is probable that other rocks of this kind will be found in the eastern part of Southern Borneo. The extent of the region in which these leucitic rocks are found, nearly 1,000 miles in length, is much greater than that of Central Italy, which is at present the best-known region of leucitic lavas.

[1] H. Bücking, *Sammlungen des geol. Reichs-Museum Leiden* (Leyden, 1902), pt. 7. 45, and 1904, pt. 8.

F. Rinne, *Zeitschr. d.d. geol. Gesellsch.*, LII (1900), 1.

[2] H. A. Brouwer, *Versl. Kon. Akad. Wetensch. Amsterdam*, 1909.

In addition to the rocks just described from Java and Celebes there are two from Sumatra that have been analyzed in order to learn whether earlier analyses of rocks from the same localities which showed relatively high alkalies were correct. One is a grayish-black pitchstone with many minute phenocrysts that occurs where the road crosses a stream in the village of Semaboer, on the south slope of Merapi volcano, in Central Sumatra. In thin section it is seen to consist of brown globulitic glass base crowded with microlites of prismoid feldspar, pyroxene, and magnetite, with scattered phenocrysts of calcic plagioclase and brown pyroxene. The chemical analysis No. 19 and the norm show that it is dacose, with normal composition for a dacitic andesite-pitchstone without visible modal quartz crystals.

The second rock from this region is dark greenish gray and aphanitic, with minute phenocrysts of feldspar and mafic minerals. It occurs in a small stream near the village of Rau Rau, at the east base of Merapi volcano. In thin section it is seen to be almost holocrystalline, with possibly a little pale-brown glass base between prismoids of plagioclase and smaller prismoids and anhedrons of pyroxene, with magnetite. There are abundant small phenocrysts of calcic plagioclase, fewer larger ones of pale augite, and numerous small olivines. The chemical analysis No. 20 and the norm show the rock is an andose with normal amounts of alkalies. It may be called an olivine-bearing pyroxene andesite.

TRIMERORHACHIS, A PERMIAN TEMNOSPONDYL AMPHIBIAN

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No genus of vertebrates occurs so abundantly in the Permian deposits of Texas and Oklahoma—it is unknown elsewhere—as *Trimerorhachis* Cope. Almost always the remains are found in bone-beds as masses, more or less cemented together, of isolated and disturbed bones, often broken, sometimes waterworn, never in anatomical relation. Isolated specimens are not often found, and in such cases where parts of a single skeleton are associated they are more or less jumbled together. The clavicular girdle only is sometimes found closely associated with the skull, filling out more or less the interval between the mandibles posteriorly. In the Chicago collections there are probably parts of at least five hundred individuals. I have counted seventy-five isolated occipital condyles, nearly as many articular ends of the mandible, and scores each of humeri, femora, ilia, scapulae, clavicles, and epipodials, for the larger part fragmentary. Very often the bones are associated with disconnected bones of *Diplocaulus;* not rarely with teeth of *Diplodus* and dipnoans; and sometimes with the smaller disconnected bones of land reptiles. All of which go to prove that the various species of *Trimerorhachis* were purely aquatic animals, living probably in shallow waters near the shores.

Different writers have expressed the opinion, and I have shared it, that *Trimerorhachis* is the most generalized of our American temnospondyl amphibians. I am now about convinced that it is the most specialized, for the following reasons:

We have very good reason to believe that this and other groups of stegocephs were, by the beginning of Permian times, already very old. The origin of terrestrial amphibians surely dates from at least as early as late Devonian times and of land reptiles from early Pennsylvanian if not Mississippian times. There can

be little doubt that *Eosauravus*, from the Coal Measures of Linton, Ohio, is a real reptile; and Watson has recently figured[1] a femur from the Lower Carboniferous of Scotland that he believes to be of a true reptile, or at least of a "precocious" amphibian. I think that he is right.

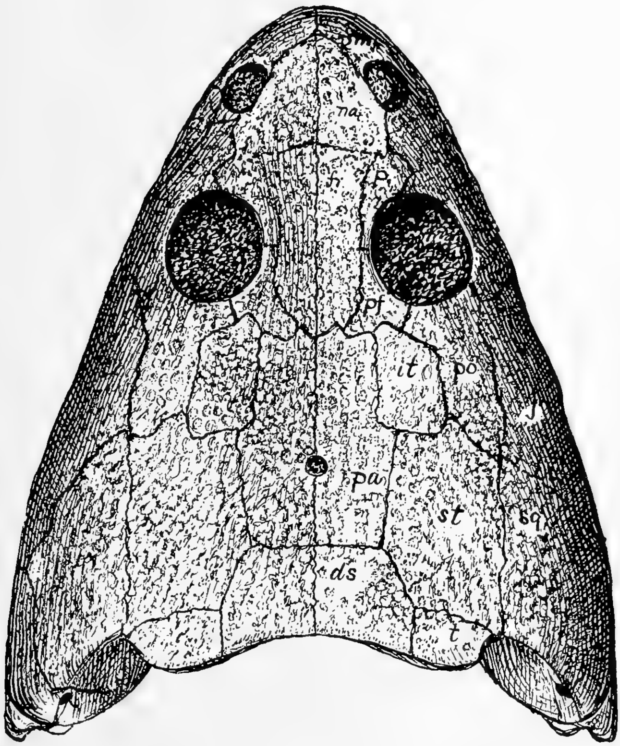

FIG. 1.—Skull of *Trimerorhachis insignis* Cope. One-half natural size: *pm*, premaxilla; *na*, nasal; *p*, prefrontal; *pf*, postfrontal; *po*, postorbital; *it*, intertemporal; *pa*, parietal; *st*, supratemporal; *j*, jugal; *sq*, squamosal; *ds*, dermosupraoccipital; *t*, tabulare; *l*, lacrimal; *m*, maxilla; *qj*, quadratojugal; *sp*, splenial; *psp*, postsplenial; *an*, angular; *art*, articular.

Trimerorhachis was a purely aquatic amphibian, as its mode of occurrence indicates and as its structure demonstrates. The characters showing aquatic adaptation are found in the anterior position of the orbits and their direction upward; in the relatively

[1] *Geological Magazine* (1914), p. 347.

small size of the limbs and the attachment of the anterior pair almost at the angle of the mandible; in the shape of the humerus, very unlike that of the terrestrial types, with the planes of its extremities but slightly divergent and with strong muscular rugosities, as in the plesiosaurs and marine turtles, situated low down; in the absence of the adductor ridge on the femur; in the relatively very short epipodials, a certain indication in crawling reptiles of swimming habits; in the conspicuous lack of ossification at the ends of the long bones; in the almost certain chondrification of the mesopodials and pubes; in the short scapulae, utterly unlike those of terrestrial amphibians; in the structure of the clavicular girdle, so like that of *Diplocaulus* as to be almost indistinguishable at first sight; and in the certainly bare skin. Further evidence is scarcely needed.

It may be granted that the first amphibians were aquatic animals, perhaps as much so as is our modern *Necturus*, although of that I am not convinced. If then, *Trimerorhachis* is a generalized amphibian, it naturally follows that it has retained its primitive structure and habits from Devonian times; that its peculiar clavicular girdle is primitive, as also the unossified mesopodials and the relatively feeble ossification of the vertebrae, the very small size of the pleurocentra, etc. It necessarily follows that if the clavicular girdle is primitive, that of the stereospondyl amphibians of the Upper Trias is also, for the buckler-like structure in these, the last of the stegocephs, is much more pronounced. On the other hand, I believe that the clavicular bones in the direct line of the ancestry of the reptiles were never large and rugose, that they began as small dermal elements and continued so in those amphibians which gave origin to the reptiles.

The occipital condyle of *Trimerorhachis* is remarkably different from that of all other known American Permian amphibians, as will be seen by reference to the figures (Fig. 5, I, J, K), in that it is single and deeply cupped, fishlike. Granted that this a primitive character, and it may be, it does not necessarily disprove the argument I make that *Trimerorhachis* was descended from terrestrial forbears, that its aquatic characters are acquired, not hereditary. Watson urges that the primitive amphibians had a closed

palate, like that of the reptiles; *Trimerorhachis* has very wide parasphenoidal vacuities, though only a slender parasphenoid. Certainly the resemblances between *Trimerorhachis* and *Diplocaulus* in the eyes, clavicular girdle, and small limbs are adaptive, not genetic, for *Diplocaulus* is a holospondylous amphibian.

In the evolution of the feet we know that aquatic adaptation has frequently resulted in the more or less complete chondrification of the mesopodials. One need only to study the progressive evolution of the mosasaurian paddle to be convinced of this.[1] And the cetaceans are still better examples.

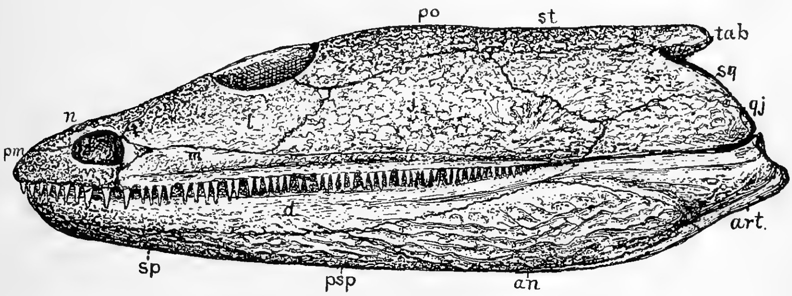

FIG. 2.—*Trimerorhachis insignis*, skull, from the side. Explanations as in Fig. 1

In the progressive adaptation of the tarsus to terrestrial life, there has been a continued loss of elements and a strengthening and closer articulation of those remaining, not only in the mammals, but in the reptiles as well. In the most primitive known tarsus, that of *Trematops*,[2] there are not less than thirteen ossified bones, four in the first row, four in the second, and five in the third. The early amphibian was a truly crawling animal, dragging its body over the surface of the ground. Its feet were directed outwardly, and the motion of the tarsus on the leg was chiefly lateral; the angle between the over-extended foot and the leg was always obtuse. In the evolution of the reptiles greater speed was attained by the elevation of the body from the ground in locomotion. No modern reptiles crawl, in the strict sense of the word, except the snakes and legless lizards—for the most part at least; many even

[1] Williston, *American Permian Vertebrates*, p. 45.

[2] Williston and Case, *Carnegie Publication, No. 181* (1913), p. 56.

rear themselves on the hind legs in running. In such locomotion the leg is brought more nearly at right angles with the plane of the plantigrade foot, and the result has been a closer union of the proximal bones of the tarsus, either with each other or with the leg bones. The earliest known reptile, from the Coal Measures of Ohio, had only two bones in the proximal row, the astragalus and calcaneum, and, even as early as the beginning of Permian times, the centralia and fifth tarsale had begun to disappear. In the elevation of the heel from the ground a further union of the proximal bones took place in some forms, *Pareiasaurus*, for instance, and some living lizards and turtles. In modern reptiles the foot is plantigrade. Among the first results of digitigradism was the closer union of the astragalus with the tibia, as seen in *Ornithomimus* and other bipedal dinosaurs, becoming more intimate in a more or less sutural union in *Rhamphorhynchus*, and the sutural obliteration in *Pteranodon* and birds, and the final union of the tarsalia with the metatarsals in the latter. In known reptiles the chief joint between the leg and foot is intratarsal, that is between the first and third rows. In the mammals it is between the tibia and the astragalus. One wonders how the change occurred in the ancestral reptiles.

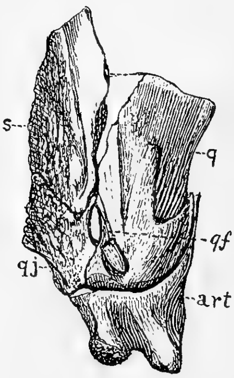

FIG. 3.—*Trimerorhachis*, part of skull, from behind: *q*, quadrate; *qf*, quadrate foramen; *art*, articular; *s*, squamosal; *qj*, quadratojugal. One-half natural size.

Not only has the vertical posture of the leg been the cause of the loss of tarsal bones, and of the change of the chief tarsal joint to the immediate end of the tibia, as in mammals, or its functional equivalent in the dinosaurs, ptderodactyls, and birds, but it has also been the cause, I believe, of the reduction of the phalanges in turtles, theriodonts, and mammals. In the most rectigrade posture of the lower leg and foot of all turtles, the land tortoises, but two phalanges remain in each toe. In the rectigrade Sauropoda no toe has, I believe, more than four phalanges; and the tarsal bones are much reduced.

Notwithstanding the abundant but disconnected material, I have been frustrated in the attempt to make out the complete anatomy of *Trimerorhachis*, and it will only be by the fortunate discovery of a connected skeleton that the tail, ribs, and feet will be made known. The figures of the skull herewith given are based chiefly upon three specimens. The most perfect of these, so far as form is concerned, is a solitary skull and clavicular girdle found by Mr. Miller in the same horizon as, and in the immediate vicinity

FIG. 4.—Clavicular girdle of *T. insignis*, from below; four-fifths natural size

of, the skeleton of *Seymouria* described by me in a previous paper. The specimen originally was perfect and undistorted, but weathering had carried away much of the thin roof, leaving the cast very smooth and complete. These parts have been completed from two other specimens of identical size and character, specimens collected in Texas nearly twenty years ago by Professor Case. The sutures separating the elements have been determined chiefly from these three specimens, aided by parts of several others, and I think can be relied upon. In their general courses and relations there is but little novelty in the figures. All of them, except the

more anterior ones, were first determined by Professor Case.[1] Dr. Huene completed our knowledge of the anterior ones, and Dr. Broom has corroborated the most of them. The most striking peculiarity of the skull is seen in the large size and posterior extension of the lacrimal.

The figure of the clavicular girdle is made from one of the numerous isolated specimens agreeing closely with that attached to the first-mentioned skull. Its comparison with that of *Cacops*[2]

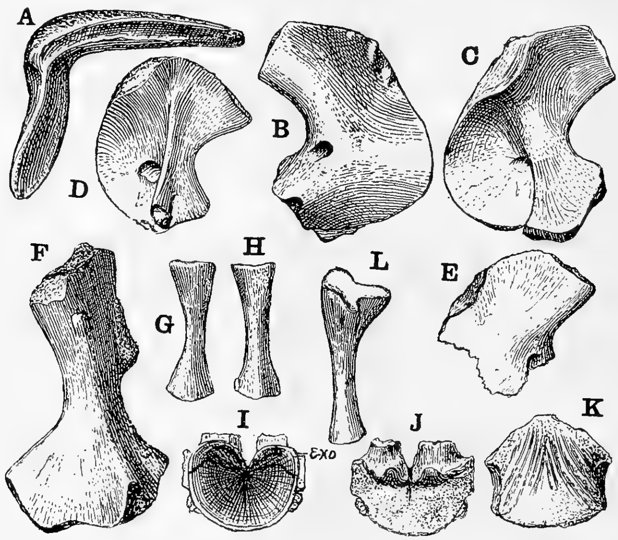

FIG. 5.—*Trimerorhachis*. *A*, right interclavicle, from behind; *B*, right scapula, from without; *C*, the same, from within; *D*, scapula of a smaller individual or species; *E*, fragment of another scapula; *F*, humerus; *G*, *H*, radii; *I*, occipital condyle from behind; *exo*, exoccipital; *J*, the same bone from in front; *K*, the same from below; *L*, ulna. All figures three-fourths natural size.

will show the extraordinary differences from the more terrestrial type of the contemporary temnospondyl amphibians. The scapula is also very peculiar in its short, rounded shape. The thickened part on its front border is apparently for articulation with the ascending process of the clavicle (Fig. 5, C). The foramen is the supraglenoid. The bone agrees closely with that figured by Case

[1] *Carnegie Publication, No. 146, (1911)* pp. 107 ff., A, B.

[2] *Bulletin American Geological Society*, XXI, Plate II, Fig. 1.

(*op. cit.*, p. 113), referred provisionally to *Zatrachys*. The coracoid has not been certainly recognized. The bone figured (Fig. 5, E, F) is probably correctly identified; possibly it is the ischium. There are no other evidences of pelvic bones, except of the ilia, of which more than fifty are in the Chicago collections. I have figured the relations of the quadrate in a specimen in which the different elements had been separated (Fig. 3). The sutural roughening for the pterygoid is very conspicuous. The separated quadrate is not very rare in collections.

Not a trace of any dermal ossicles has been detected in any of the abundant material; the skin unquestionably was bare. Certain skin ossifications or calcifications, however, have been detected in a few specimens, first mentioned by Cope and later by Case. They consist of very thin sheets, like leaves of thin paper, probably of calcified cartilage, overlying each other, a dozen or more, in places. They probably sheathed the under side of the abdomen.[1]

The bones figured were all found dissociated, and probably belong to different species, though there is not much difference in size. A number of species of *Trimerorhachis* have been described, though it is a futile and almost hopeless task to identify them. I assume that the skull and pectoral girdle belong to *T. insignis*, because they agree, so far as I can determine, with the species so named. I cannot forbear entering a protest here against the heedless naming of species of fossil reptiles, and especially of the Permian vertebrates, based upon fragmentary material or, much worse, upon difference in size. Such names seldom advance science. Nor does anyone care much, at the present time, about species; they are, for the most part, a nuisance. Specific determinations will become of use only when precise differences

[1] Rather singularly dermal ossicles have very rarely been observed in the air-breathing vertebrates of the American Permian—so far as I am aware only in *Eryops*, though ventral ribs or scales are a rather common characteristic of the reptiles. Very recently I have observed in *Pantylus*, the strange reptile of which only the skull has been known hitherto, a continuous covering extending probably over the whole body, though possibly only on the under side, composed of a mosaic of small, smooth, bony scutes two or three millimeters in diameter. The vertebrae and ribs show as great peculiarities as does the skull.

are required for geological correlations, or for minor problems in evolution, and their naming should, so far as possible, be deferred until much more material is available for comparison. As I have said elsewhere, the genus is practically our unit for most of the older vertebrates. In very few of the Permian vertebrates do we yet know what specific differences really are; we have few measuring sticks yet to measure them by; and it is only the morphological characters of the genus and higher divisions that concern us much in paleontology. Perhaps this protest is a sort of belated

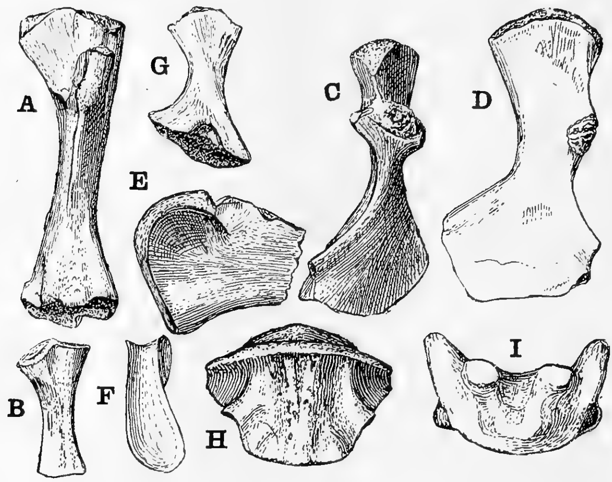

FIG. 6.—*Trimerorhachis*. *A*, left femur, from behind; *B*, tibia; *C*, humerus, radial side; *D*, another humerus, dorsal side; *E*, coracoid (?); *F*, the same, end view; *G*, right ilium, outer side; *H*, anterior intercentrum, from below; *I*, the same, from behind. All figures three-fourths natural size.

repentance on the part of one who has himself contributed hundreds of specific names to zoölogical and paleontological literature, but the many hours he has spent in the endeavor to identify organisms from vague and imperfect descriptions at least give him the right to urge temperance in the making of more or less worthless types! About twenty-five specific names were given to the mosasaurs of Kansas. The most diligent search of abundant material has resulted in the recognition of scarcely a half-dozen. The same fate probably awaits the great majority of the specific names which have been given to the Permian vertebrates.

There are various species among the material known as *Trimerorhachis;* of that there can be no question; as also of *Eryops;* but when one species is based upon an isolated scapula, for instance, another on the comparative size or minor distinctions of a skull, and a third upon a vertebra, one is tempted to ask: What is the use?

It need not be said that the differences of occipital condyle, pectoral girdle, vertebrae, and limbs in *Trimerorhachis* are of more than family importance. Doubtless some day, when we know the forms better, the genus or genera will be separated into a group of higher rank.

THE GEOLOGY OF LIMESTONE MOUNTAIN AND SHERMAN HILL IN HOUGHTON COUNTY, MICHIGAN[1]

E. C. CASE and W. I. ROBINSON
University of Michigan

Limestone Mountain and Sherman Hill consist of three small residual masses of dolomite in southeastern Houghton County, Michigan. The first is partly divided and is locally known as Big and Little Limestone. These paleozoic outliers have been the subject of some previous study, but the visits of the geologists have been brief and the work incomplete. During the summer of 1913 the authors spent six weeks in the region in an attempt to determine the exact age and structure of the beds. The result of this work will be published in full by the Michigan Geological Survey, but the results obtained contain so much of interest that a brief preliminary report seems desirable.

Limestone Mountain lies half a mile east of the little station of Hazel on the Mass City branch of the Mineral Range Railroad and directly north of the track. Sherman Hill lies one and a half miles northeast of Limestone Mountain.

The fossils collected were determined by the junior author and finally submitted to Dr. E. O. Ulrich of the United States Geological Survey for revision and for the determination of the exact horizons. For this and for many helpful suggestions we desire at this point to express our thanks to Dr. Ulrich.

As finally determined, the stratigraphy of the beds is as follows:

Mid-Devonian.—All that is known of this horizon is a single mass of chert protruding from the talus on the southeastern slope of Big Limestone. It yielded four fossils:

Chonetes coronatus var.
Productella cf. *navicella* and *spinulicosta*
Spirifer aff. *pennatus*
Cystodictya cf. *hamiltonensis*

[1] Published by permission of the director of the Michigan Geological Survey.

These fossils are few in number but so characteristic that there can be no doubt of the presence of mid-Devonian rocks in or near Limestone Mountain.

Niagaran (*Lockport*).—A bed of very siliceous material was found on the south slope of Big Limestone. It could not be traced for any very great distance nor could any satisfactory determination of its position be made. No trace of a similar layer was found in any other outcrop. The fossils collected are:

Streptelasma spongaxis
Zaphrentis stokesi
Duncanella(?) sp.
Clorinda cf. *ventricosa*
Pentamerus sp.
Chonchidium decussatum(?)
Dalmanella cf. *elegantula*
Leperdita aff. *cylindrica*
Loxonema sp.

Middle to Upper Richmond.—Three fossils were found with the Niagaran material which show the presence of this horizon, but the bed could not be distinguished stratigraphically from the Niagaran.

Favosites asper
Columnaria alveolata
Plectorthis whitfieldi

Upper part of the Lower Richmond (*Arnheim*).—Near the southeastern corner of Big Limestone a thin layer of dolomite was found in the bottom of a deep gully. The layer was almost perpendicular, but as it was in the zone of broken talus we cannot be sure that it was in place.

Crinoid columnals
Coeloclema oweni
Mitoclema minutum
Mesotrypa patella
Bythopora striata
Rhynchotrema perlamellosa
Rhynchotrema capax
Conularia formosa
Primitia cincinnatiensis
Tetradella persulcata var.

Ceratopsis robusta
Calymene, a new species allied to *C. fayettensis* and *C. mamillatus.*
Conodont

Lower Richmond.—A bed about ten feet thick occurs at the top of Little Limestone and on the southeastern slope of Big Limestone which contains fossils of this age. The matrix is siliceous and the fossils are, as a rule, silicified.

Streptelasma rusticum(?)
Halysites gracilis
Iocrinus aff. *I. crassus*
Orbiculoidea (*?Schizotreta*), n. sp.
Rafinesquina, n. sp.
Leptaena unicostata, two new varieties
Plectambonites sp.
Plectorthis whitfieldi
Plectorthis kankakensis
Dalmanella aff. *rogata*
Dinorthis subquadrata
Hebertella, n. sp. aff. *H. insculpta*, *H. fausta*
Platystrophia sp.
Rhynchotrema capax

Galena (*Stewartville or Upper Galena*).—Sixty feet of heavy-bedded, cream-colored dolomite beneath the Lower Richmond on Little Limestone and at the top of Sherman Hill. The fossils occur near the top of the bed.

Cyrtolites cf, *retrorsus*
Liospira cf. *angustata*
Hormotoma(?) major
Lophospira minnesotensis
Maclurea crassa
Maclurina manitobensis
Maclurina cuneata
Trochonema umbilicatum
Fusispira subbrevis
Spiroceras sp.
Salpingostoma cf. *expansa Hall* and *buelli Whitfield*

Decorah (*Upper Blue*).—Below the last-described layer there is an old quarry on the eastern face of Little Limestone in a thin-bedded dolomite. The top layers are gray, the lower layers cream-colored with blotches of dark-red iron stain. Identical fossils

occur in both layers. A layer similar to the lower one, occurring on Sherman Hill, also carries Decorah fossils but shades up into the heavy dolomite without the intervening thin-bedded layer. The fossils found in the quarry are:

Crinoid columnals
Ceramophylla frondosa
Trematopora(?) primigenia
Halloporina crenulata
Arthrostylus sp.
Arthroclema sp.
Rhinidictya mutabilis
Rhinidictya fidelis(?)
Arthropora simplex
Escharopora subrecta
Escharopora confluens
Orthis tricenaria
Streptelasma profundum
Strophomena incurvata
Strophomena septata
Dalmanella rogata(?)
Hormotoma salteri canadensis
Aparchites sp.

Upper Black River (*Upper Bluff*).—At the extreme southwestern part of Big Limestone and eight feet above the sandstone there occurs a heavy-bedded, cream-colored, fossiliferous layer of dolomite. The beds above and below are completely barren, so far as we could determine.

Ctenodonta nasuta
Ctenodonta gibberula
Endodesma(?), n. sp.
Cyrtodonta billingsi
Cyrtodonta cf. *billingsi*, n. sp.
Cyrtodonta cf. *huronensis* and *subcarinata*
Cyrtodonta cf. *tenella*
Cyrtodonta, n. sp.
Vanuxemia aff. *niota Hall* and *subrotunda Ulrich*
Vanuxemia sp.

Potsdam (*Jacobsville*).—This is the lowest horizon in all the outliers and the only stratum which was observed at all three localities. It is a dull-brown, coarse, poorly cemented sandstone with

occasional streaks of a very fine conglomerate or a very coarse sandstone. On Big Limestone and Little Limestone the pebbles of the conglomerate are of quartz and about the size of a pea; near Sherman Hill the pebbles are larger and there is much chert and greenstone. No fossils were found in this layer. It has been referred by Lane to the Jacobsville Sandstone, probably of Potsdam age.

The hills are broken in a very intricate manner by minor faults of comparatively recent age. They are, in many cases at least, due to under cutting of the dolomite and to slumping. There are some major faults, notably the one between Big and Little Limestone, which involve the sandstone below. Unfortunately the data available do not serve to determine the age of the great Keweenawan fault; the most that we can say is that there was serious disturbance of the region at least as late as after mid-Devonian time.

The most important point brought out by this study is the demonstration that the paleozoic seas of the times mentioned above extended well into, if not over, the Northern Peninsula of Michigan, and our paleogeographic maps must be so far altered as to include that region within the areas of deposition. Dr. Ulrich has called attention to the similarity which exists between the Pentameroid forms of Big Limestone and those of the Far West indicating a broad extension of the Niagaran sea in that direction.

In conclusion the senior author desires to state that most of the field work and the determination of the fossils was done by the junior author and that to him in a large measure is due the credit for the recognition of the extension of the paleozoic sea over the Northern Peninsula of Michigan.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA FOR 1909, 1910, 1911, AND PART OF 1912

EDWARD STEIDTMANN
University of Wisconsin

III. LAKE SUPERIOR REGION AND ISOLATED PRE-CAMBRIAN AREAS OF THE MISSISSIPPI VALLEY

Adams[1] states that the formations of the Cuyuna iron-bearing district of the north-central part of Minnesota consists of a series of complexly folded Upper Huronian slates and other clastics interstratified with lenses of iron formation, and intruded by basic Keweenawan rocks; the whole trunkated and almost completely covered by patches of Cretaceous conglomerate and a thick mantle of glacial drift, presenting a surface of gentle relief.

The iron formation occurs as a series of parallel and overlapping lenses or belts, generally associated with magnetic attraction on the limbs of trunkated anticlines, trending northeast-southwest. It presents all gradations between original cherty iron carbonates, and its secondary phases, ferruginous cherts, hard, soft, low-grade hematite ores, and amphibole magnetite rocks.

Adams follows Van Hise and Leith in his belief that the ores occur in places which have been favorable to the circulation of surface solutions, such as in inclined basins formed by pitching folds superimposed on the major anticlines or by the intersection of dikes with the walls adjacent to the iron formation lenses. Since the ores in such localities grade into ferruginous cherts, and cherty iron carbonates where the circulation of solutions was less favorable, he concludes that the ores have resulted from the oxidation of the cherty iron carbonates to ferruginous cherts and from the latter by the leaching of the silica. The amphibole magnetite rocks probably have developed from the anamorphism of cherty iron carbonates.

[1] Francis S. Adams, "The Iron Formation of the Cuyuna Range," *Econ. Geol.*, V, No. 8 (1910), 729–40; *ibid.*, VI, No. 1 (1911), 60–70; *ibid.*, VI, No. 2 (1911), 156–80.

Allen[1] reports that the stratigraphic succession in the Iron River district of Michigan from the Brule river northward consists of fine-grained, ellipsoidal basalts, and green schists probably Keewatin; Lower Huronian cherty and slaty dolomite, quartzite, and slates; Upper Huronian slates and graywackes, containing iron formation lenses in at least four horizons, and associated with basic igneous rocks, mainly extrusive. Unconformities between the major units are inferred but not known, owing to the drift-covered condition of the contacts.

The rocks are complexly folded, but the details of the structure are obscure. The major axis of folding is northwest-southeast in the eastern part of the district, east and west in the central portion, and southwest to northeast, farther west.

The Upper Huronian is divisible into three main belts or stratigraphic horizons running parallel to the major structure, that near the Brule River on the south being the lowest. The southern belt consists dominantly of slates and iron formation lenses, the productive portion of the district. The next overlying belt is ill defined, consisting dominantly of extrusive greenstones. The uppermost belt, lying farthest north, consists of slates, graywackes, basic extrusives, and iron formations.

The iron formations comprise slaty and cherty iron carbonates, ferruginous cherts and slates, magnetic, chloritic, sideritic slates, altered greenalite, and secondary soft, hydrated hematite ores. The ore bodies occur mainly as irregular sheetlike masses parallel to the bedding or as pockets related to structural basins. They are bounded by phases of lean iron formation, extrusive basic rocks, or carbonaceous shale, or any combination of these.

Buckley[2] confirms Haworth's conclusions that the pre-Cambrian rocks of Missouri consist of contemporaneous granites and rhyolites intruded by diabase dikes. About 200 feet of slate and conglomerate are exposed on Pilot Knob. The igneous rocks are correlated by Buckley with the Laurentian and the sediments with the Huronian, but the reasons for this correlation are not evident.

[1] R. C. Allen, "The Iron River Iron-bearing District of Michigan," *Mich. Geol. and Biol. Surv.*, 1910, 144 pp., 17 pls., 18 figs.

[2] E. R. Buckley, "Geology of the Disseminated Lead Deposits of St. François and Washington Counties, Missouri," *Missouri Bur. Geol. and Mines*, IX (1908), Pt. 1, 259 pp., 39 pls., 10 figs.

Crane[1] describes the pre-Cambrian iron ores of Iron Mountain, Missouri, as veins of specular hematite containing apatite and tremolite, which occur in porphyry; and detrital ores derived from the vein ores. The ores of Shepherd Mountain near by are also vein ores, but their principal constituent is non-titaniferous magnetite, with pyrite and a little clay. The Pilot Knob ores are soft blue, banded hematites, locally with ripple marks, and rest conformably on a nearly smooth floor of porphyry, and grade laterally into a porphyry breccia. The formation is separated into two parts by a claylike bed. The ore above the claylike bed is more distinctly stratified and grades upward into a hundred feet of breccia.

Crane concludes that these ores represent replaced tuffs, from the fact that they grade into breccias both along the strike and the dip. It has been found extremely difficult to explain the origin of the Pilot Knob ores under any hypothesis offered thus far. This is due in large part to the fact that only a small remnant of the ores, and the rocks associated with them have been left by erosion. The principal objection to the replacement hypothesis is found in the interstratification of non-iron-bearing materials with the ores. Under the replacement hypothesis it would be necessary to assume that the replacement of certain layers was accomplished, while others remained unaltered.

Crane does not correlate the pre-Cambrian rocks of Missouri with certain Lake Superior pre-Cambrian rocks, as some of his predecessors have done.

Grout[2] infers from analyses of Keweenawan diabase in different stages of alteration that the fresh rocks contain more copper than the altered, the copper being in the form of a basic silicate in the fresh rock, and that the more basic diabases originally contain more copper than the more acid types.

Grout[3] finds that chemical and mineralogical analyses of the Keweenawan lavas of the Kettle River region of Minnesota indicate

[1] G. W. Crane, "Iron Ores," *Missouri Bur. Geol. and Mines*, Vol. X, (1912) pp. 36–39, 107–44.

[2] Frank F. Grout, "Keweenawan Copper Deposits," V, No. 5 (1910), 471–76.

[3] F. F. Grout, "Contribution to the Petrography of the Keweenawan. *Jour. Geol.*, XVIII, No. 7 (1910), 633–57.

a close magmatic relationship between them and ascribes the differences to differentiation.

Analyses of fresh lavas containing 0.01 to 0.03 per cent of copper are cited by him as proof that the Keweenawan ores were derived from the lavas. This conclusion is open to the objection that a similar content of copper has been found in many rocks where no ores were developed.

Lane[1] finds that the waters from the deep levels of Michigan iron mines are high in sulphates and chlorides which are associated with calcium rather than sodium. The carbonate radical is low and decreases with depth. Their silica content is like that of ordinary cold surface waters, and they contain almost no iron.

A. C. Lane and A. E. Seaman[2] divide the pre-Ordovician of Michigan into Keewatin, Huronian, and Cambrian or Primordial. The Keewatin corresponds to the Archaean; the Huronian to the Algonkian; and the Cambrian or Primordial to the Keweenawan and Potsdam of Van Hise and Leith. Lane believes that the Potsdam and Keweenawan should not be placed in different geologic periods because of their stratigraphic and structural similarity and continuity, and because no unconformity can be found between them.

Lane[3] recognizes the following groups of Upper Michigan Keweenawan rocks from the base upward: (1) Bohemian Range group: consisting mainly of basic lavas but with intrusive and and effusive felsites and coarse labradorite porphorites; also intrusive diabase dikes and gabbro aplites, with a total thickness of about 9,500 feet; (2) the central mine group: mainly lavas of an augitic ophite type with infrequent sediments; this includes the Allouez conglomerate, the Calumet and Hecla conglomerate, and the Kearsarge lode; the total thickness is placed at from 3,823 to 25,000 feet; (3) the Ashbed group: consisting of basic lavas of the

[1] A. C. Lane, "Michigan Iron Mines and Their Mine Waters," *Jour. Can. Min. Inst.*, XII (1909), 114–29.

[2] Notes on the Geological Section of Michigan," part of *Annual Report* for 1908, Michigan Geological Survey, 1909, 120 pp.

[3] A. C. Lane, "The Keweenawan Series of Michigan," *Mich. Geol. and Biol. Surv.*, 1909, 2 vols., 932 pp.

Ashbed type with scoriaceous sediments and a little conglomerate. The thickness may be between 1,456 and 2,400 feet; (4) the Eagle River group: a group of basic lava flows with frequent beds of sediments; its thickness is between 1,417 and 2,300 feet; (5) the Great Copper Harbor conglomerate: a coarse, heavy conglomerate from 1,800 to 2,200 feet thick; (6) the Lake Shore traps: a series of thin flows having a total thickness of about 800 feet; (7) the Outer Copper Harbor conglomerate: from 1,000 to 5,000 feet in thickness; (8) the Nonesuch shales: black, fine-grained, micaceous sands and grits, about 500 feet in thickness; (9) the Freda sandstones: red, impure sands, shales, and conglomerates; 4,000 or more feet in thickness.

The Keweenaw rocks dip at a high angle toward the north under Lake Superior and in general strike parallel to the axis of the Keweenawan Peninsula. They have also been disordered by numerous faults. The great Keweenaw fault, a longitudinal fault dipping toward Lake Superior and cutting across the bedding, is regarded by Lane as a thrust fault due either to the contraction of the earth's crust or to uplift produced by the injection of a large sill. Movement along this fault began before Cambrian time and has taken place since the deposition of the Niagara limestone. There are also other faults both parallel to the bedding and longitudinal faults which cut across the bedding. There are numerous vertical or nearly vertical transverse faults which have produced an offset of the beds to the right in going toward the northeast.

The Keweenaw rocks of northern Michigan as well as those of other parts of the Lake Superior region show strong evidence of consanguinity. Thus in all of these rocks potassa appears to be subordinate to soda except in extremely siliceous varieties, and in general the potass is very low. Free quartz is not abundant. There is never an excess of alumina which would result in the development of corundum nor are there any ultra-alkaline or ultra-basic rocks. The content of iron is decidedly high. According to Iddings' classification, the commonest type would be auverngnose. No rocks are known which contain less of both silica and alkalies.

Lane finds that the composition of these rocks is on the sodic side of a line which he regards as a hypothetical eutectic between

alkalies and silica. In harmony with this, he finds very few cases of porphyritic texture and very little difference in the time of crystallization of the various constituents. In general, labradorite has come out first and augite last in the dikes and flows, while in the deep-seated intrusives, the order is reversed. He also regards the development of olivine rather than enstatite or hypersthene as in harmony with the position of the analyses with respect to his hypothetical eutectic.

The igneous rocks are all crystalline. If any glass has existed in these rocks it has long since become devitrified. The following varieties of texture are described by Lane: (1) ophitic, in which the augite crystals occur as a cement, enclosing idiomorphic feldspars; this texture characterizes the central portion of the basic effusives; (2) doleritic texture, in which the feldspar is much coarser than the augite; (3) glomeroporphyritic or navitic, in which the feldspars present a great range of size and tend to aggregate in bunches; (4) porphyritic hiatal, characterized by coarse feldspar phenocrysts imbedded in a ground mass in which the feldspars are very small; (5) amygdaloidal texture in which blow holes have resulted from the escape of gases; (6) microlitic, a texture which develops around amygdules; it is characterized by fine slender prisms of feldspars; (7) vitrophyric, consisting of a glasslike ground mass in which a few crystals are imbedded; it is always found within a few millimeters from the margin; (8) the graphic texture, consisting of intergrowths of quartz and feldspar; it is very common in the interstices of the Bessemer gabbro and is also very abundant in felsite conglomerate pebbles; (9) spherulitic texture, found only in salic rocks; it is characterized by small spherules of rock; (10) mozaic texture, consisting of equigranular quartz and feldspar.

The factors which determine the grain of igneous rocks are stated by Lane to be the chemical composition, temperature, and pressure of the magma. He presents a mathematical discussion of the relation between grain and cooling.

The mean annual temperature of the air on Keweenaw Point is between 38° and 42° Fahrenheit. The temperature of the upper mine levels at a depth of from 100 to 236 feet varied from 43° to 50°, and the average was nearly 43°. The rate of increase of

temperature with depth is found to be very small, about 1° in 103 feet. Lane suggests that this low temperature gradient may be due to chemical reactions which absorb heat, such as the deposition of copper; the diffusivity of the strata, permitting the early and free escape of heat; downward absorption of water carrying with them cooler temperatures of the surface; recent deposition of surface drift, and the relative exhaustion of the internal supply of heat by the Keweenaw and earlier eruptions.

The mine waters of the region are of three types: (1) normal surface carbonate waters, in which calcium and magnesium are in excess of sodium; (2) waters of intermediate levels, 1,000–2,000 feet, high in sodium and chlorine, in which lime and magnesium are subordinate; (3) waters of deep levels in which lime chloride dominates, and in which magnesium is absent or nearly absent and sodium subordinate. Waters of type (2) are associated with the richest part of the lodes and with the occurrence of silver. The deep waters are regarded by Lane as connate waters, but not necessarily derived from the ocean of that time.

Lane regards the lavas and sediments as the original source of the copper. All the lavas and sediments contain at least 0.02 per cent or more of primary copper. The lavas and sediments he believes were deposited rapidly one upon the other, the lavas retaining their heat for a long period of time. Water and gases containing chlorine were given off by these lavas and these caused a hydration of the silicates of the rock, developing chlorite and epidote and some native copper. This first alteration, however, he believes was insufficient to account for the development of all the copper, since the occurrence of the copper in the porous parts of the formation indicates that circulation of water must have taken place. The circulation which Lane believes developed the ores was set up largely through the cooling of the rocks. As the rocks cooled they contracted and afforded openings for water which was sucked in from the surface. A large part of this water entered into chemical combinations with the rocks and thus a very strong and long-continued downward circulation must have been set up, since the greater portion of the rocks have been hydrated to chlorite. This reaction also involved solution of sodium silicate, copper, and lime

in the form of chloride. The precipitation of the copper Lane believes took place according to the experimental results of Fernekes, the copper being taken out of solution by means of ferrous chloride in the presence of alkalies, principally in the form of calcite and prehnite and datolite, since the copper shows most intimate associations with these minerals.

Paige[1] states that the pre-Cambrian rocks of the Llano-Burnett region consist of a sedimentary series of gneisses and schists and banded iron formation lenses composed of alternating layers of magnetite, quartz, and silicates, the whole intruded by a large development of granites and minor basic igneous rocks.

Richardson[2] states that the lowest pre-Cambrian rocks of the El Paso district are white and red quartzites, fine, round grained, intruded by diabase dikes. About 1,800 feet are exposed in the Franklin Mountains. The dip of the beds is 20°–45° west. They are separated by a slight unconformity from 1,500 feet of massive red rhyolite porphyry overlying them. The rhyolite is separated by a marked unconformity from the Cambrian beds above it.

Sharwood[3] publishes a large number of analyses of the rocks, minerals, ores, and waters of the Homestake mine, as a supplement to *Professional Paper No. 20* of the U.S. Geological Survey.

Steidtmann[4] finds that a minority of the joints of the Baraboo quartzite are clearly related to the local folding; namely, joints parallel to the bedding, and strike joints which dissect the bedding. The abundance of the joints is related to the intensity of the folding. The majority of the joints are vertical, gaping, continuous cracks, which are not related to the local folding, but probably belong to larger units of structure.

Todd[5] states that the pre-Cambrian rocks have been penetrated by borings in the Aberdeen-Redfield quadrangles of South Dakota.

[1] Sidney Paige, "The Mineral Resources of the Llano-Burnett Region, Texas," *Bull. 450, U.S. Geological Survey*, 1911, pp. 103, maps and illustrations.

[2] G. B. Richardson, "The El Paso Quadrangle," *U.S. Geological Survey, Folio 166*, 1909, 11 pp., 2 pls., maps, etc.

[3] W. J. Sharwood, "Analyses of Some Rocks and Minerals from the Homestake Mine, Lead, South Dakota," *Econ. Geol.*, VI, No. 8, pp. 729–89.

[4] "The Secondary Structures of the Baraboo Quartzite," *Jour Geol.*, XVIII, No. 3 (1910), 259–70.

[5] J. E. Todd, "The Aberdeen Redfield Folio," *U.S. Geological Survey, Folio 165*, 1909, 13 pp.

They include a quartzite, believed by Todd to be the Sioux quartzite, granite, and mica schist.

F. T. Thwaites[1] concludes that the sandstones of the Wisconsin coast of Lake Superior form a single conformable series, which may be separated into two groups, which grade into each other, viz., a lower group, consisting of steeply tilted red feldspathic sandstones and fragments of igneous rocks, shales, and conglomerates; and an upper group, composed of slightly disturbed red and white quartz sandstones. The contact of the upper group with the Middle Keweenawan traps is marked by a thrust fault. A small amount of conglomerate in the sandstones along this contact may indicate a local unconformity. Since the faulting, folding, and erosion of the traps and the deposition of the sands went on simultaneously, there is no reason for believing that this conglomerate represents a great time interval, if indeed any. Irving placed the upper group in the Cambrian since he regarded the sandstone faulted against the Middle Keweenawan traps at St. Croix Falls and the same in age as the sandstone at the falls of the Amnicon River. Thwaites shows that the former is Upper Cambrian, but that the latter is older.

The conditions of deposition of the sandstones appear to have been dominantly subaerial as indicated by current marks, mud cracks, rain prints, irregular and curved bedding, depressions in shale beds filled with sands, red color, and feldspathic content.

The relations of these sandstones to the known Cambrian of Wisconsin is still doubtful. Thwaites points out certain differences between them. The Cambrian is marine as shown by fossils, and its sands generally consist of rounded, yellow quartz grains often enlarged so as to show crystal faces, and cemented dominantly with calcite. The sandstones of the Lake Superior coast lack fossils and were largely deposited under subaerial conditions. Their feldspathic content, angular grain, irregular quartz enlargements, red color, irregular and curved bedding, and lack of calcite cement contrasts strikingly with the characteristics of the known Cambrian.

Van Hise and Leith's[2] "The Geology of the Lake Superior Region" is a summary of the geology of the Lake Superior region.

[1] F. T. Thwaites, "Sandstones of the Wisconsin Coast of Lake Superior," *Wis. Geol. and Nat. Hist. Surv. Bull. 25*, 1912, pp. 109.

[2] "The Geology of the Lake Superior Region," *U.S. Geological Survey, Monograph 52*, 1911, 641 pp., 49 pls., 76 figs.

Their conclusions on the origin of the sedimentary iron formations from which the ores were derived by secondary concentration will be considered here. The unaltered iron formations consist essentially of alternating bands of chert, iron carbonate, oölitic greenalite, and iron oxides, varying in thickness from a few feet to more than a thousand feet. It is inferred that they are chemical sediments from their composition, lack of fragmental texture, and bedding. They show a great variety of lithologic associations, subaqueous ellipsoidal greenstones, tuffs, acid extrusives, and normal sediments, viz., quartzites, limestones, and slates.

From the fact that the iron content of the thick formations and the sediments associated with them is greater than the iron content of the rocks which antedate them, and since normal weathering tends to fix iron in place rather than to cause its transportation and deposition elsewhere on a large scale, it is concluded that the iron in these thick iron formations is not the result of a normal cycle of erosion, but was contributed from some unusual source. Further evidence for this conclusion consists in the fact that their thickness, composition, and structural characteristics do not resemble either the bog or the glauconite deposits of the present time. Furthermore, their calcium-magnesium ratio is the reverse of that prevailing in normal sediments. On the other hand, the small bodies of iron formation, particularly those interbedded with slates, are similar in their composition, thickness, and association to deposits formed by the solution, transportation, and deposition of iron by processes of normal weathering. The unusual source of iron as well as of the silica of the thick deposits Van Hise and Leith believe may have been from the submarine extrusions of basic lavas, with which the iron formations are more or less associated in time and in place. They postulate that the lavas may have contributed the iron solutions directly, or that iron solutions may have been formed from the interaction of the hot lava and sea water, and that the lavas may have furnished iron solutions by weathering.

They show that salt water acting on hot basalt forms sodium silicate, and that sodium silicate in the presence of iron salts will form oölitic iron silicate and silica which are thrown down in alternating bands, and that when carbonic acid gas is present, iron carbonate

may form instead of the silicate, while in the presence of abundant oxygen, the development of alternating bands of ferric oxide and silica may take place. Thus by simple laboratory reactions, the composition, textures, and structures of the Lake Superior iron formations are duplicated.

The monograph cites a number of cases of the intimate association of sedimentary iron formations of various ages with eruptive rocks in places outside of the Lake Superior region, but fails to name one of the most striking cases which presents a close parallel to that of the Lake Superior region, namely, the oölitic, banded, marine Devonian hematites of the Harz Mountains, which are intimately associated with subaqueous basalt extrusions, diabase, tuffs, and limestones.

Wright[1] discusses the relations of the ophites, oligoclase gabbro and aplite of Mount Bohemia at Lac La Belle on Keweenaw Point.

The oligoclase gabbro is intrusive into the ophite flows, but the aplite inclosed by the gabbro is regarded as a differentiate from the gabbro. The process of differentiation is supposed to have been accomplished by fractional crystallization, convection currents, and general upward movement of the lava. The intrusives contain veins of sulphides and arsenides of copper which are regarded as genetically related to the native copper deposits in higher horizons of the ophites. The contact metamorphism of the ophites by the gabbro is slight. Diopside has changed to uralitic hornblende, and there is a slight increase in silica and oxidation of iron at the contact.

[1] F. E. Wright, "The Intrusive Rocks of Mount Bohemia," part of *Ann. Rept.* for 1909, Michigan Geological Survey, 32 pp.

[*To be continued*]

DISCOVERY OF THE NORMANSKILL GRAPTOLITE FAUNA IN THE ATHENS SHALE OF SOUTH-WESTERN VIRGINIA

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During the last year, as opportunity was afforded, the writer has had under investigation a Cambro-Ordovician section in southwestern Virginia, not only as a contribution to our knowledge of the geology of this section of Virginia, but more especially for the purpose of correlation, if possible, with the standard section north. In connection with this work the discovery was made that the Athens shale abounds in graptolites. The determination of some thirty or more species and varieties shows that the majority, if not all, are the same as those which characterize the Normanskill shale of New York.

As the paper on the entire Cambro-Ordovician section will not appear for some months, this occurrence is thought of sufficient moment to warrant the publication of this note in advance.

A number of papers have appeared in recent years discussing the Normanskill graptolite fauna and the correlation of the shales containing it. The literature on the subject shows that the formation has, from time to time, been referred to every possible position from the Hudson River formation to the Chazy. In 1901, however, Dr. Ruedemann showed[1] that the slate which contains this fauna in the Hudson River region underlies the Utica slate, and correlated it with the lower or middle Trenton. This correlation, based almost entirely on paleontologic evidence, is in general accordance with the views held by R. R. Gurley, H. M. Ami, and Charles Lapworth. Later, in 1903, Weller found the fauna in New Jersey, and, from his study of the beds containing it, concluded that they are about equivalent to the middle portion of the

[1] *Bull. 42, N.Y. State Mus.*

typical Trenton limestone of New York, or even to a position below the middle.

Dr. Emmons long ago expressed the belief that the formation extended through to Virginia, but very little is reported of it from the south. Professor E. A. Smith, state geologist of Alabama, reports a small fauna of about eight species occurring in calcareous shale associated with Trenton limestone, which he believes to be upper Trenton. This find, however, could demonstrate nothing more than the general fact of the Trenton age of the Normanskill fauna. Stratigraphic proof of the age of the formation has not been found in New York for the reasons stated by Ruedemann, that the whole mass of rocks, which contains Trenton, Utica, and Lorraine beds in similar lithologic development, is in New York as well as in Canada cut off from the neighboring terranes by extensive faults, thus apparently frustrating all attempts at a stratigraphic solution of the problem. As the primary object of this preliminary note, in addition to calling attention to the many species of graptolites occurring here, is definitely to locate on stratigraphic grounds the exact position of the formation containing this fauna, a description of the relevant portion of the section is given.

The Cambro-Ordovician section under investigation is in Roanoke County, Virginia, about 11 miles north of Salem, and extends from the top of Catawba Mountain northward across Catawba Valley to the foot of North Mountain, a distance of about 6,000 feet. This section is characterized by the unusual occurrence of all the strata from top to bottom in regular chronologic order, there being no faults or folds to disturb the natural sequence. Furthermore, the strata are most favorably disposed for observation, dipping at an angle of 45° on Catawba Mountain, and gradually rising to 70° in the valley. The new state road just being graded down the side of the mountain and across the valley exposes every stratum in the section from the top of the Medina on the south face of Catawba Mountain to the base of the Cambrian at the foot of North Mountain.

At the latter point there is a fault bringing the Carboniferous in conjunction with the Cambrian, which terminates the section. The section therefore includes every stratum from the top of the

Medina to the base of the Cambrian, with each in practically normal development. The section embraces the following members:

Medina sandstone, gray sandstone, quartzitic	400–500 feet
Sevier shales, calcareous shales and sandstones, and bluish-gray impure limestone. All very fossiliferous	1,200 feet
Utica shale, fossiliferous blue to black shales, and limestones with shale partings	300 feet
Tellico sandstone, red and gray sandstones	400 feet
Trenton, dark gray-blue, and black limestones with shale partings	500 feet
Athens shale, black carbonaceous shales, carrying graptolites, trilobites lingulas	560 feet
Black River, coarse dark coralline limestone with bands of smoky marble in upper part	250 feet
Birdseye (Lowville), pure dove limestone, with Tetradium	50 feet
Chazy, cherty magnesian limestone, with Maclurea magna	500 feet
Beekmantown, pure and impure, cherty and magnesian limestone	1,700 feet

As the Ordovician is now known to extend as far south as Alabama, evidenced by the finding of Ordovician graptolites by Professor E. A. Smith, and as the writer has found Ordovician fossils in the black slates in eastern Virginia, and as the Ordovician is known to occur far to the north and west of this point, it is but fair to presume that the sediments represented in the rocks of this section were laid down far within the confines of that ancient sea and that the section is therefore in normal development. Furthermore, as will presently be shown, the fossils—graptolites and others—occurring here are the same as those found in equivalent strata in New York and Canada, and Dr. Ruedemann makes the statement that the essential identity of the Alabama fauna with the Normanskill graptolite fauna cannot be gainsaid. It therefore seems most probable that the sea was open and continuous from Alabama to Canada, and that sedimentation was virtually the same throughout. The section here should, therefore, be practically the same as in New York. A notable difference occurs in the Tellico sandstone, which, however, may be but an expansion of the Dolgeville shale, a name proposed by Cushing in 1909 for the shaly phase of the upper Trenton which he previously described as "Trenton Utica passage beds." It may therefore be stated with greater

confidence that the Catawba section is normal and in general accord with the section north.

The Utica consists of two parts. The upper is slaty or shaly and more or less calcareous, carrying several species of corals and brachiopods, while the lower part is slaty limestone and limestone beds with some shale partings. The thin beds, which vary in thickness from one to four inches, are largely made up of brachiopods. The Tellico beneath is composed of red and gray sandstone and shale, with a massive, hard, quartzitic, ferruginous sandstone 30 feet thick at its base. It carries in addition some kaolin and flakes of mica. This basal member rests immediately upon the Trenton. The formation is about 300 feet thick. The Trenton limestone is conformable with the Tellico above, the two dipping at an angle of 60° S. The contact between them is sharp and clear. The limestone ceases suddenly as the sandstone comes in. The upper part of the Trenton consists of rather massive, dark, blue-gray limestone with some thin shale partings. The lower part is made up of alternating layers of solid black homogeneous limestone and thin bands of slate, the limestone bands varying in thickness from two to four to six inches. Fossils have not been observed except near the contact with the Athens shale, where a number of small trilobites were found. These probably belong to the Athens shale as the same species have since been found in the lower part of it. The Trenton formation is 500 feet thick, and passes into the Athens shale without perceptible break. Fresh-rock contact, however, has not been observed. On the surface there is frequently a slight depression due to differential weathering marking the contact. The Athens formation is compact, black, carbonaceous shale, slightly calcareous. The lower part contains some layers which are massive, and which resemble the Trenton limestone; but on examination the lime content is found to be no greater than in the more shaly portion above. The massive character is found to be due to less pressure having been applied to this part of the formation. It contains innumerable forms of *Dicellograptus sextans* which are not at all crushed or flattened, while the graptolites in the upper part of the formation are more or less flattened. These forms look as if they could be taken out bodily and sections of them be made for

microscopic study. The fossils show a smooth, polished, black, chitinous surface. Near the middle of the formation along with the various species of graptolites there are many forms of *Triarthrus becki*, and *Leptobolus walcotti.*

In New York the formation—Normanskill—carrying the same fossils is described by Ruedemann as a deep bluish-black, thick-bedded argillite with conchoidal fracture, and iron-stained cleavage. This description applies equally well to the Athens shale of Virginia. In *Folio 16 U.S.G.S.* Keith says of the Athens shale in Tennessee: "It is everywhere composed of blue and black shales, which do not vary in appearance. The black shales are found at the bottom of the series and contain lingula and numerous graptolites. The blue shales gradually replace the black shales in passing up through the series, and when fresh consist of thin light blue, shaly limestones." It is evident that the blue shales above the black, which he says consist of thin, light-blue, shaly limestone, represent the Trenton in that section, and the black shales below with the included graptolites and lingulae represent the Athens. Keith, in that section, also represents the Athens shale as immediately underlying the Tellico sandstone. Immediately below the Athens shale he describes a limestone formation consisting of massive blue and gray limestones, shaly and argillaceous limestones, and marbles. These beds are all very fossiliferous, and fragments of corals, crinoids, brachiopods, and gastropods are so abundant as sometimes to make most of the bulk of the rock. The upper beds of the formation consist of more or less coarsely crystalline marble. This is exactly the position of the marble beds in the Black River formation, and anyone at all familiar with it will see at once from the above description that the Athens shale of Tennessee, as well as that in Virginia, rests upon the Black River.

The conglomerate R. Ruedemann described[1] as occurring in the Normanskill of New York, although looked for, was not observed. Neither does the Rysedorph conglomerate occur here, which Ruedemann describes as intercalated in undoubted Normanskill, and containing in pebbles as well as matrix, as its youngest fossils, those of lower Trenton aspect. There does occur here, however,

[1] *Bull. 42, N.Y.S.M.*

a conglomerate at the base of the Lowville (Birdseye), separating it from the Chazy, which is in many respects like the Rysedorph conglomerate, but not in fossil content. While not essential to the points involved in this note, yet, for the sake of clearness, a brief description of it will be given when that point is reached in describing the section.

The mass of the Athens shale (Normanskill) dips at an angle of 70° S., and is about 550 feet thick. It occupies the center of Catawba Valley, which structurally is anticlinal, and through which meanders Catawba Creek. The shales where exposed in bluffs weather into thinly cleaved plates or leaves of a dark peppery gray on which the graptolites are not readily observed. The completely weathered soil is ocherous yellow, due to the presence of iron in the original rock. The fossils are in places preserved in pyrite, and as a rule where the seams open upon weathering the surface is rusty. On such surfaces the black fossils stand out clearly. The formation maintains its general character throughout, and terminates abruptly against the underlying Black River limestone, and does not here rest upon lower Trenton limestone. In this respect its position differs from the Normanskill of New York, for Ruedemann states in his summary that these shales rest on lower Trenton limestones. Mather, however,[1] in describing the Black River, in which is included the Lowville, several times states that the Black River is underlaid by a brecciated limestone and overlaid by slates. It would appear, therefore, that in certain localities, at least in New York, the succession of strata for this part of the Ordovician is the same as in Virginia; for these Black River limestones and slates of which Mather writes occur at Mount Moreno, and Mount Moreno is one of the four typical localities where complete or nearly complete Normanskill faunas have been collected. The conclusion to be drawn from the above is that the Normanskill there rests on Black River limestone as in Virginia.

The question has also arisen whether the Normanskill is a distinct development of the Ordovician or only a clastic phase of the Trenton. For this locality at least, it would appear to represent a distinct epoch in the Ordovician; for the formation is at least

[1] *Nat. Hist. N.Y.*, Part IV, "Geology," pp. 405-6.

500 feet thick, which is as thick as the Trenton above, and maintains its individual character throughout, in respect to lithology as well as fossil content. While no decided break or hiatus has been observed between it and the Trenton, yet there is a marked change in sedimentation. The shale deposit ceases and limestone appears, and along with it an entire change of life. Between it and the formation below the change in sedimentation is still more marked, for the fine black sediment of the Athens shale rests immediately upon the thoroughly crystalline limestone and marble of the Black River. The feature that would most closely associate the Athens shale with the Trenton is the occurrence of *Triarthrus becki* in the shale, but while it has been reported from the Trenton of other localities, it has not been observed to occur in the Trenton at this point. The Trenton here is singularly devoid of fossils. Everything considered, conditions here point to the same conclusion for the Athens shale as those reached by Lapworth, Ami, and Ruedemann for the Normanskill of New York, viz., that it represents a distinct epoch of the Ordovician.

The species of graptolites and other forms from the Athens shale thus far determined are as follows:

Climacograptus bicornis Hall
Climacograptus modestus
Climacograptus parvus Hall
Climacograptus putillus Hall, mut. eximus
Climacograptus putillus Hall
Climacograptus scalaris
Climacograptus typicalis Hall, mut. spinifer
Climacograptus scharenbergi Lapworth
Corynoides gracilis Hop., mut. perungulatus
Cryptograptus tricornis Carruthers
Dicellograptus sextans Hall
Dicellograptus intortus Lapworth
Dicranograptus spinifer Lapworth
Dicranograptus nicholsoni Hopkins
Dicranograptus ramosus Hall
Didymograptus serratulus Hall
Didymograptus sagittacaulis Gurley
Didymograptus subtenuis Hall
Diplograptus amplexicaulis Hall
Diplograptus angustifolius Hall

Diplograptus foliaceus var. incisus
Diplograptus foliaceus Murchison, var. acutus Lapworth
Diplograptus foliaceus
Dictyonema ———
Glossograptus quadrimucronatus Hall, var. cornutus
Lasiograptus mucronatus Hall
Leptobolus walcotti
Nemagraptus exilis Lapworth
Nemagraptus exilis Lapworth, var. linearis
Nemagraptus gracilis Hall
Phycograptus laevis (Graptolithus laevis Hall)
Retiograptus eucharis Hall
Triarthrus becki

An interesting feature of a number of these species is their occurrence in original colonies, or synrhabdosomes. Several species show synrhabdosomes which have not heretofore been observed. All of the above listed forms occur in fresh material, are very distinct, and apparently not at all distorted, or flattened. The graptolites occur in all parts of the formation from top to bottom.

The Black River formation beneath the Athens shale is made up of dark to black coarsely crystalline limestone and marble, 250 feet thick. It contains several distinct bands of marble between the middle and the top. One occurs at the top. The marble is of a smoky-gray color, rather coarsely crystallized, and fossiliferous. The bed near the middle is about 6 feet thick, the other beds somewhat less. The mass as a whole, except the lower part, appears to be a coral reef on which flourished crinoids, bryozoa, brachiopods, gastropods, and, in the lower part, orthoceratites. The lower part of the formation is more or less siliceous and cherty, which on weathering produces knots over the surface. The Black River was laid down on an unevenly eroded surface of dove-colored limestone, the Lowville (Birdseye). The hiatus here shown is well marked. A sharp line can be drawn between the coarse, dark, impure limestone above and the pure fine-grained, dove-colored limestone beneath. The Lowville contains *Tetradium cellulosum* in abundance, and near the bottom large, coiled shells. The formation is about 50 feet thick and terminates in a very coarse, brecciated limestone conglomerate, which separates it from the Chazy below.

This appears to be a basal conglomerate. It is composed of fragments ranging in size from an eighth of an inch to pieces 6×15 inches. They consist of angular grains and pebbles of quartz, pieces of dark chert 4×6 inches, large blocks of magnesian limestone, and angular fragments of all sizes from the underlying cherty magnesian limestone of the Chazy and Beekmantown formations. The fossils, among them *Ophileta*, are imbedded in such a manner with the fragments as to indicate that they were fossils when inclosed. The conglomerate is from 6 to 8 feet thick. This evidently is the conglomerate referred to by Mather as occurring at Mount Moreno. It is in all probability the basal conglomerate described by Weller from New Jersey. The writer has noted its occurrence in Maryland and at several places here in southwestern Virginia. It has been observed here at two points, in the same stratigraphic position, 12 miles apart at right angles to the strike, showing that the sea was encroaching upon the land, beating down the ancient Chazy-Beekmantown limestone cliffs, and scattering the fragments along the shore from Canada to Tennessee[1] as the sea advanced.

Enough of the section has now been described to locate definitely the stratigraphic position of the shales carrying the Normanskill graptolite fauna. It is evident that the position assigned them here, below the Trenton and above the Black River, is correct for Virginia and Tennessee, and is in harmony with the statement of Mather for part of New York. It is also in general accord with the views entertained by Ami, Lapworth, and Gurley, who, on paleontologic evidence only, thought the fauna should be placed in lower Trenton if not below the Trenton.

Dr. Ruedemann[2] admits the essential identity of the Alabama fauna with the Normanskill, but says it is open to question whether this upper Trenton of Alabama is exactly equivalent to that of New York. In the light of the manifestly clear section in southwestern Virginia, it is difficult to see how these same shales could be upper Trenton in Alabama. Furthermore, the section as given

[1] Keith describes a similar conglomerate "of unknown age" from that state, which, from his description (*U.S. Folio 16*), is in all probability the same.

[2] *Mem. II*, *N.Y.S.M.*, p. 12.

by Professor Smith reads as if the strata were very much faulted, folded, or otherwise disturbed. The graptolite beds are seen to be alternating with beds of chert and sandstone, and they are stated to occur at the top of the series, not far below chert beds with sub-Carboniferous fossils. The latter indicates either profound thrust faulting or an interval of non-deposition.

While the evidence gathered by those who, in the North, have been wrestling with the problem of the age and stratigraphic position of the Normanskill graptolite fauna points to nothing more definite than probably lower Trenton, and no further extension south than the present site of New Jersey in the Appalachian geosyncline, it is believed that the facts here presented definitely fix both age and position, and at the same time show a southern extension as far as Tennessee.

SUMMARY

The great number of graptolites occurring in the Athens shale of Virginia, and the fact that nearly all of those thus far identified are Normanskill forms, leaves no doubt as to the identity of this fauna with that of the Normanskill. The Athens shale carrying this fauna is to be correlated with the Normanskill shale of New York.

The stratigraphic position of the Athens shale in southwestern Virginia is clearly below the Trenton and above the Black River, and it seems probable that when discovered at points intermediate it will be found to occupy the same position.

Reviews

The Antiquity of Man in Europe. By James Geikie. New York: D. VanNostrand Co., 1914. Pp. 305. Illustrated.

As stated in the preface, this volume consists of a series of ten lectures delivered to a mixed audience and hence contains much elementary matter. It takes up first the migrations of the southern and temperate, the tundra or snow-loving, and the steppe fauna and flora in their relation to climatic changes in the Pleistocene epoch. The second and third lectures discuss cave deposits and human and animal relics and the archaeological stages of culture. The fourth lecture deals with river deposits and the succession of archaeological culture stages contained in them. A brief discussion of the loess and its relation to tundra and steppe conditions is also presented.

In chap. v, under the heading "Glacial Action," there is not only a discussion of glacial action proper, but also of the formation of screes, of the flowing soils of artic regions, and of the breccias of Gibraltar. Erosion of rock basins and overdeepening of valleys by glacial action is given prominence. In the following chapter, which deals with the glaciation of northern Europe, submarine basins in the Irish Sea and near the Hebrides, and the overdeepened valleys and fiords of Norway are referred to glacial action. Rock rubble in non-glaciated areas is also described. The thickness, extent, and direction of flow of the British as well as Scandinavian ice fields form the main theme. In chap. vii, the glaciation of the Alps and overdeepening of Alpine valleys and the glaciation of other mountains in middle and southern Europe are discussed; also the general climatic conditions of middle Europe in glacial and interglacial stages, and conditions favorable for loess deposition. The climatic conditions of the several interglacial stages and the different kinds of interglacial deposits are treated in the next chapter. The two concluding chapters deal with the history of the Pleistocene epoch, and the relation of the archaeological culture stages to the general glacial and interglacial stages. Estimates of geological time are briefly considered.

The volume is illustrated by 21 full-page plates of which 17 show characteristic animals, plants, and Paleolithic implements. The

remainder are photographs of Alpine scenes. There are also four small folded maps as follows: A. Europe during the second glacial epoch; B. Europe in interglacial times; C. Europe during the third glacial epoch; D. Europe during the fourth glacial epoch. The fourth glacial epoch as interpreted by Geikie corresponds to only the early part of the fourth or Würm Stage of Alpine glaciation. The fifth and sixth glacial stages of Geikie's classification seem likely to be represented in America by such strong readvances of the ice as the readvance to the Port Huron morainic system.

The aim of the author has been to deal with the question of the antiquity of man in Europe from the geological standpoint, and he maintains that it is chiefly by following geological methods of investigation that the successive stages of human culture are put on a firm and reliable basis. The discussion is restricted wholly to Europe with no reference to Asiatic invasions. In fact no intimation is given that the present people in Europe had any other line of descent than through the Paleolithic and Neolithic man. The earliest Paleolithic man is, on geological grounds, estimated to have appeared somewhere between 250,000 and 500,000 years ago. A more precise estimate is not considered possible in the present state of knowledge.

Frank Leverett

Ann Arbor, Michigan

Twenty-second Annual Report of the Bureau of Mines, Ontario. Vol. XXII, Part I, 1913. By Thomas W. Gibson. Pp. 284, pls. 84, sketch maps 5, sheet maps 4.

This bulletin contains ten articles and reports. It commences with a review of the mining operations and production of the province during the year 1912. The value of the mineral output was $48,341,612—15 per cent greater than that of 1911. The decreasing production of the Cobalt region was compensated by a great enough rise in the price of silver to show an increase in the total value produced during the year. The Porcupine region added a million and three-quarters dollars' worth of gold to the annual output of the province. The year showed an increase in the output of both metallic and non-metallic minerals, but a much greater increase in the metallic.

The mines of Ontario are described by Mr. E. T. Corkell. He gives details of mine development, treatment of the ores, and production of the mines.

The geology of Whiskey Lake and the Massey Copper Mine area is reported upon by Dr. A. P. Coleman. The former workings were made first for copper and later for gold. The latter district has been worked for copper. Neither of the districts is commercially productive.

With the addition of the District of Patricia, the province of Ontario was given a location for a railway terminus at the mouth of Nelson River within territory formerly included in Manitoba. The location and surveying of this section are reported by Mr. J. B. Tyrrell. He describes his reconnaissance observations upon the agricultural and geological features of the parts of Patricia which he visited. In the northern part, the ancient crystalline rocks are overlain by fossil-bearing Ordovician and Silurian rocks. The Labradorean ice sheet deposited enough of a till covering upon the bared rocks of the northern part of the region to offer considerable agricultural possibilities. There are many instances of rock scorings of two glacial sheets. These are the Patrician and the Labradorean ice sheets. In Trout Lake region there is one set of striations which runs north 25° to north 40° west, planed and grooved by a glacier which moved south 40° west. The first striae were made by the Patrician glacier, which had its center in the highland southeast of Trout Lake; the later glaciation from the northeast was that of the Labradorean ice sheet. Upon the Hayes River, Mr. Tyrrell found evidence of the westward movement of the Patrician glacier followed by a southeastward advance of the Keewatin glacier. He considers, therefore, that the Patrician ice sheet preceded both the Labradorean and the Keewatin invasions in this region.

Mr. A. L. Parsons deals with the Lake of the Woods and other mineral areas in northwestern Ontario. He gives a detailed description of many of the ancient rocks of the region. A brecciated contact between granite bosses and Keewatin rock is common. The explanation offered is that the granite was intruded between the highly schistose layers of the metamorphosed Keewatin rocks. In most of the massive Keewatin rocks this contact breccia gives place to coarser, banded structure.

The West Shining Tree Gold area is described by Mr. R. B. Stewart. The rocks are chiefly of Keewatin age. The gold content is in quartz veins. The veins are very irregular in size and persistence; they are generally less than 4–6 feet wide and break into stringers. Many of the veins contain visible gold. A large amount of gold has been deposited along the fracture lines of the quartz, and a little in the bordering schists. The district is only in the very first stages of development.

"Glacial Phenomena of Toronto and Vicinity," by Dr. A. P. Coleman, and "Moraines North of Toronto," by Mr. Frank B. Taylor, are articles dealing with the glacial developments near the city of Toronto. The most interesting feature of the region is the Toronto interglacial formation which includes the Scarboro and the Don beds. The Don beds furnish many fossils of plants and shellfish. Among vertebrate fossils the remains of a bear, a bison, and two kinds of deer have been recognized. The whole assemblage of plants and animals implies a warmer climate than is there at present; one is implied comparable to the present climate of Ohio or Pennsylvania. The Scarboro beds have yielded fossils of 72 species of beetles and 3 mosses; these remains betoken a cool climate. During the Don interglacial period climatic conditions must have been such as to preclude the presence of an ice sheet within hundreds of miles of Toronto.

T. T. Q.

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1. *Glaciation of the Puget Sound Region.* By J. Harlen Bretz. Washington Geol. Survey Bull. 8. 1913. Pp. 244; pls. 24, figs. 27, maps 3.
2. *Bibliography of Washington Geology and Geography.* By Gretchen O'Donnell. Washington Geol. Survey Bull. 12. 1913. Pp. 64.
3. *Geology and Ore Deposits of the Covada Mining District.* By Charles E. Weaver. Washington Geol. Survey Bull. 16. 1913. Pp. 87, pls. 5, figs. 3.

1. The area studied stretches from the Canadian border 170 miles south to the divide between the Chehalis and Columbia rivers, and from the base of the Cascade Mountains to the Olympics, about 50 miles east and west. This region has been studied by several other geologists; but hitherto the glacial geology has not been so thoroughly treated as in this report. It is a detailed study of the region which discusses the Pleistocene history and later diastrophism. Terminal moraines, recessional moraines, extra-morainic outwash, the formation of glacial lakes, and drainage changes are the principal topics discussed in connection with the drift.

The submarine features are as remarkable as the surficial deposits. The region is characterized by deep marine channels or troughs. The greatest trough, that of Puget Sound, contains Admiralty Inlet, 60 miles

long and from 3 to 4 miles wide. For a distance of 45 miles its depth is between 575 and 900 feet below sea-level. The origin of the troughs is a subject which still requires further careful work. The author does not accept glacial erosion, drift deposition, or preglacial and interglacial stream erosion as the sole agents of the trough-making, but advances the opinion that each one, and in some cases all, of these have been causes of some of the troughs.

In Pleistocene times there were two glacial epochs, the Admiralty and the Vashon, with an interglacial epoch known as the Puyallup. The glacial invasions were both from the north. In the interglacial epoch there was an uplift of the region about 1,000 feet above the present level which was followed by stream erosion to submaturity stage. In postglacial times there has been submergence of 250 to 280 feet below present level and re-emergence.

2. The bibliography mentioned above is an expansion of the bibliography of Washington geology published by the Survey in 1910, and includes all publications up to date with an introduction of geographical material. It is provided with a full subject index.

3. The mining district covered by the third report is about 30 miles northwest from Spokane. The ore is argentiferous galena in quartz veins. Pay ore has been found in few places, and the output is very small.

T. T. Q.

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1. *Report of the Topographic and Geologic Survey Commission of Pennsylvania*, 1910–1912. By RICHARD R. HISE. Pp. 102, pls. 21, figs. 23, maps 5.
2. *Graphite Deposits of Pennsylvania.* By BENJAMIN L. MILLER. Topographic and Geologic Survey of Pennsylvania. Report 6. 1912. Pp. 143, pls. 17, map 1.

1. The first of these reports includes a bibliography of the publications of the state and of the United States Geological Survey relating to Pennsylvania. Other appendices are: "Preliminary Report on the York Valley Limestone Belt," by M. L. Jandorf; "Geological Origin of the Freshwater Fauna in Pennsylvania," by Dr. A. E. Ortmann; "A Peridotite Dike in Fayette and Greene Counties," by Lloyd B. Smith; "The Mineral Production of Pennsylvania," by R. R. Hise.

2. The report on graphite includes a general discussion of the history, properties, occurrence, and origin of graphite, with a full statement of its

distribution in the United States. It is concerned chiefly with the production, methods of mining, and manner of milling of graphite in Pennsylvania.

T. T. Q.

The Geography and Industries of Wisconsin. By RAY HUGHES WHITBECK. Wisconsin Geol. and Nat. Hist. Survey Bull. No. 26. 1913. Pp. 94, pls. 20, figs. 48.

This bulletin is essentially a geographical treatise on the natural resources of the state. It is designed primarily for use in the schools. The mineral production, forest industries, agriculture, manufacturing, and transportation are all in turn discussed with relation to indigenous opportunities and necessities. The volume is well calculated to meet the needs of those for whom it was written.

T. T. Q.

Krystallisationskraft. By RAPHAEL ED. LIESEGANG. Naturwissenschaftliche Umschau, No. 12, Beilage der Chemiker-Zeitung, 1913, Nos. 154, 155.

This article contains a brief summary of the work of various writers who have contributed, since 1836, opinions or experimental data on this subject. Mention is made of the work of Bruhns and Mecklenburg, who used potassium nitrate and alum solutions and reached the conclusion that capillarity and adsorption were probably more important factors than the force of the growing crystals. They found under conditions of alternate wetting and drying that the movement due to capillarity was actually complete before crystallization took place.

The author does not believe that the force of growing crystals can be considered a factor in geologic processes.

E. A. S.

Useful Minerals of the United States. By SAMUEL SANFORD and RALPH W. STONE. U.S. Geol. Surv. Bull. No. 585. Pp. 250. Washington, 1914.

This bulletin contains alphabetical lists of the useful minerals occurring in each of the states, with the localities in each state where each is found. A glossary and mineral index of thirty pages is appended.

A. D. B.

Mineral Resources of Southwestern Oregon. By J. S. DILLER. U.S. Geol. Surv. Bull. No. 546. Pp. 146, figs. 23, plates 10.

After a brief introductory chapter describing the general topographic and geologic features of southwestern Oregon, the writer proceeds to describe the gold-quartz lode mines, copper mines and prospects, placer mines, and the platinum, mercury, nickel, and coal resources of the region.

A. D. B.

Oil and Gas Fields in Wayne and McCreary Counties, Kentucky. By M. J. MUNN. U.S. Geol. Surv. Bull. No. 579. Pp. 105, figs. 6, plates 6. Washington, 1914.

About a third of the bulletin is devoted to the stratigraphy and larger structural features—the remainder is concerned with the history of production of oil, stratigraphy of the oil sands, and the details of the structure in the oil fields within the region. Structure contours are used to good advantage in the discussion.

A. D. B.

Handbuch der Mineral Chemie, Vol. II. By C. DOELTER. Pp. 848+xvi, figs. 37, plates 3. Dresden and Leipzig, 1912–14.

Proceeding with the detailed discussion of the chemistry of minerals (see review of Vol. I, *Jour. Geol.*, XXI, 465–67) this volume is concerned with quartz and the silicates. Practically all of the experimental work of recent years is included in a condensed and thoroughly digested form, and this volume should be exceptionally valuable as the chemistry of the silicates is only vaguely understood at present, and active research is being prosecuted along this line. Numerous tables of physical and chemical data are included, making reference to the original literature unnecessary in most cases. The detailed description of methods of synthesis which characterized Vol. I is continued throughout this volume, making it of great value as a reference work to the chemist as well as to the mineralogist.

A. D. B.

Ore Deposits in the Sawtooth Quadrangle, Blaine and Custer Counties, Idaho. By J. B. UMPLEBY. U.S. Geol. Surv. Bull. No. 580-K. Pp. 221–49, plates 4.

A report of the geology and ore deposits, based on a twelve days' reconnaissance trip. Silver-lead, silver, zinc, and gold veins have been worked more or less in the region but present production is negligible.

A. D. B.

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THE EVOLUTION OF HERBACEOUS PLANTS AND ITS BEARING ON CERTAIN PROBLEMS OF GEOLOGY AND CLIMATOLOGY

EDMUND W. SINNOTT AND IRVING W. BAILEY
Bussey Institution, Harvard University

Evidence from living and fossil plants has often proved of value in geological investigations. A large part of our information as to the climate of the various regions of the earth during ancient times, particularly in the Mesozoic and Tertiary, has been derived from a study of the plants composing the various fossil floras with reference to the climatic conditions under which their modern representatives live. A gradual differentiation and refrigeration of climate in the north and south temperate zones since the close of the Cretaceous have been pretty clearly indicated by such investigations.[1] Botanical evidence, particularly that derived from a study of the distribution of living and fossil plants, has also been of value to the student of ancient geography in providing support for such theories as that of a closer connection between North America and eastern Asia just before the Glacial period; a recent elevation of the coastal bench in eastern North America; and a more or less intimate union between Australia, New Zealand, and South America in recent times by means of an antarctic continent or archipelago.

[1] There is, of course, ample evidence that similar periods of refrigeration have occurred earlier in the earth's history, notably in Cambrian and Permian times.

In fact, the builders of all the various hypothetical land bridges have used an abundance of phytogeographical data.

Botanical evidence which has heretofore been of use to the geologist, however, has been mainly derived from a study of *flora* rather than of *vegetation;* a study of the species, genera, and families which inhabited a region in past or present time and not of the various plant types which they represent. It is now possible, however, to supplement this evidence by an investigation of the habit of growth of the various elements in a flora, with especial reference to the relative proportions of herbs and woody plants, particularly among dicotyledonous angiosperms.[1] Data thus obtained derive their importance from the following facts which recent investigation seems pretty thoroughly to have established: (1) The earliest angiosperms were woody plants. (2) Herbs attained but very little prominence until the beginning of the Tertiary, since which time they have increased very greatly in number. (3) The gradual refrigeration of the climate of the temperate zones during the Tertiary, with the appearance of a well-marked winter season, seems to have been the factor responsible for the development of most herbs, which are plants well able to withstand cold winters, in the form of seeds or underground roots and stems. (4) This herbaceous vegetation, which reached its greatest development in the land mass of the north temperate zone, was thus composed of plants which were very hardy, agressive, and rapidly dispersed, and it consequently spread far and wide into other and warmer portions of the globe.

The evidence on which these conclusions are based may be summarized briefly as follows:

From paleobotany.—The vast majority of angiospermous remains, especially from the Cretaceous, are of plants whose nearest modern relatives are always trees or shrubs. In the Tertiary, especially the middle and latter parts of the period, remains of herbaceous plants are much more frequent. This geological evidence,

[1] The monocotyledons were apparently derived in very ancient times from the primitive dicotyledonous stock in adaptation to an aquatic habitat. They are almost entirely herbaceous, and their few woody forms are evidently recent rather than primitive.

however, cannot be regarded as entirely conclusive, since the delicate leaves of herbs are probably less readily preserved than are the tougher ones of most trees and shrubs. Among the lower orders, however, it is quite certain that the ancient fossil lepidodendrids, sigillarians, and calamites, for example, were woody plants, and that their modern representatives are all herbaceous.

From anatomy.—The most important anatomical difference between a woody plant and an herb lies in the degree of activity of the cambium. In the former group this tissue lays down a thick ring of wood which increases in size year by year. In the latter the woody ring is either very much thinner or, more frequently, is broken up into separate bundles. In such cases the so-called "interfascicular" cambium, opposite the gaps between the bundles, is much reduced or quite inactive; and in the bundles themselves the cambium is often merely vestigial. That such a discontinuous cambium is not the primitive type is shown by the fact that in various ancient groups which were just acquiring secondary growth (as shown by the structure of their fossils) the cambium always began as a perfectly continuous ring and did not arise from the union of isolated fascicular cambia.

Furthermore, in the vascular tissue of most herbs, medullary rays and wood parenchyma are poorly developed or absent. These structures, however, always characterize the more primitive types of angiosperm wood.

These two pieces of evidence point to the conclusion that the herbaceous stem has been reduced from a primitive woody one. This reduction has been accomplished by a marked decrease in the activity of the cambium, and in many instances also by an increase in the height and width of certain of the medullary rays, resulting in the breaking-up of the continuous woody ring into a series of bundles. In its general topography the stem of an herb resembles that of the young twig of its woody relatives.

From phylogeny.—It is generally admitted that the angiosperms have been derived either from the Bennettitales or from some stock derived from the Coniferales. There are no herbaceous forms in either of these groups. Among living angiosperms it is uncertain whether the nearly naked flowered Amentiferae or the complete

flowered Ranales and their allies are the most ancient members of the phylum, but the distinction in all probability belongs to one of these two groups. The Amentiferae are almost exclusively woody plants, as are the vast majority of the Ranales.

Evidence that herbaceous angiosperms are of comparatively recent origin is also presented by a study of families which possess both woody and herbaceous members and in which it is possible to determine, on floral or other evidence, the relative antiquity of the two types. In the Leguminosae, for example, the Mimosoideae and Caesalpineae are with very few exceptions trees or shrubs; and the Papilionatae, which seem clearly to be more recent than the other two groups, contain almost all the herbs in the family. In a number of other orders and families (Umbelliflorae, Violaceae, Polemoniaceae, Borraginaceae, etc.) and even in certain genera, notably *Potentilla,* the same fact may be observed. In practically every case where the phylogeny of a mixed group can be definitely established it is found that the woody members are more ancient in type than are the herbaceous ones.

Still further evidence pointing to the same conclusion is furnished by a general survey of the distribution of herbs among the families of the dicotyledons. Of the 240 families belonging to this group of plants listed in Engler's *Syllabus* (7th ed.), 121, or a trifle over 50 per cent, are entirely woody, whereas only 35, or 14 per cent, are entirely herbaceous. Eighty-four families, or 35 per cent, possess both woody and herbaceous forms. Of these 14 are rarely herbaceous and 18 rarely woody. Of the entirely herbaceous families, almost all are either parasitic, insectivorous, or water plants, or are monotypic, and hence can lay no strong claim to primitiveness. Practically all typical land herbs belong to families which have woody members, as well; but many more than half of the families possessing woody species (121 out of 205) include no herbaceous forms at all. Had the original type of angiospermous vegetation been herbaceous, it would be very improbable that over half of all the families should have lost their herbaceous members. If woody plants are primitive, however, we can easily see that herbs might have arisen in only about half of the families. These facts are especially significant

because in number of *species* herbs are fully equal to woody plants.

From phytogeography.—That the differentiation and refrigeration of climate in the temperate zones, particularly in the north temperate land mass, have been the chief factors in the development of herbs is indicated by the present distribution of these plants over the earth. In temperate regions, which are subject to winter temperatures considerably below freezing, from 75 to 85 per cent of the dicotyledonous plants are herbaceous in habit. In arctic and alpine regions, which are still colder, from 85 to 90 per cent or more of the dicotyledons are herbs. The relative proportion of the two types is precisely reversed in tropical countries and under climates where freezing never occurs. In such regions only from 25 (or less) to 40 per cent of the dicotyledons are herbs. Table I presents an analysis of the dicotyledonous vegetation in several typical temperate, arctic, alpine, and tropical regions.[1]

TABLE I

Region	Number of Species	Number of Herbs	Percentage of Herbs
Northeastern United States	2,280	1,748	76
Rocky Mountains	2,206	1,910	86
Ellesmereland	76	71	93
Great Britain	927	821	88
Germany	1,117	947	84
Switzerland	1,899	1,726	90
Iceland	221	200	90
Brazil	15,981	4,092	25
British West Indies	2,249	675	30
Tropical Africa	8,577	3,560	41
Java	3,188	867	27

This evident adaptation of herbs to life in regions subject to low temperatures at certain seasons makes it very probable that the advent of a cold winter has indeed been the chief factor responsible for their origin, for there is a fairly constant relation between the minimum winter temperature of a region and the proportion of

[1] In all of the tropical regions of which the floras are tabulated here, there are more or less extensive upland or mountain areas which possess a relatively temperate climate and hence a fairly high proportion of herbs. In the tropical lowlands in each case, however, only from 10 to 15 per cent of the dicotyledons are typically herbaceous.

woody plants in its flora. The advantages of the herbaceous habit in a temperate climate are obvious. The plant is able to complete its whole life-cycle during the summer season and to live through the cold winter either under ground or in the form of seeds. Plants which cannot resist cold are thus able to thrive in temperate regions. It is a noteworthy fact that families whose members have aerial stems which are generally able to withstand low temperatures—such as the Fagaceae, Betulaceae, Salicaceae, Caprifoliaceae, and, in fact, most of the woody families of temperate regions—have few or no herbaceous species, for they have been well able to get along without them.

The gradual refrigeration of climate probably first dwarfed and stunted the primitive arborescent vegetation, killing outright many of the more delicate types, such as the fig, laurel, and cinnamon (of which we find fossil remains almost within the arctic circle); and then, by continually killing back the whole year's growth of these stunted shrubs, it probably converted them gradually into perennial herbs. The annual herb, which starts from seed each year, is evidently a still more recent development.

The herbaceous type thus developed has usually proved very vigorous and adaptable. It is far superior to the woody one in the relative amount of seed which it produces; and the fact that its life-cycle from seed to seed is completed in only one or two seasons, rather than in the long period of years which is necessary with most trees and shrubs, allows it to become dispersed much more rapidly. Its ability to live through adverse conditions of drought, in the same way that it does through those of cold, gives it an advantage in dry regions, and many herbs seem to have arisen in these dry regions in adaptation to discontinuity in moisture alone.

It seems most probable that herbs have evolved in the temperate portions of the Southern Hemisphere and in mountainous, desert, and even tropical habitats all over the world. The majority of the important herbaceous genera, however, seem from their present distribution to have originated in the temperate land mass of the Northern Hemisphere. The ease of dispersal over wide areas and the consequently keen competition among a great variety of plants have resulted here in the development of an exceedingly hardy and

aggressive herbaceous flora. Most of the dominant and widely spreading plants of today, including practically all weeds, originally belonged to this northern herbaceous vegetation. These "Scandinavian" plants have taken advantage of every opportunity to become widely dispersed and to establish themselves in other parts of the world, with the result that in the temperate regions of the Southern Hemisphere, for example, there are well over two hundred genera and even fifty or more species which are identical with northern ones; and in practically all the south temperate floras more than half of the herbaceous genera either have their center of distribution in the Northern Hemisphere or are very common there. There are comparatively few southern genera, on the contrary, which have successfully invaded the North.

That the evolution of herbs has taken place for the most part since the close of the Mesozoic is indicated not alone by the fact that refrigeration of climate seems to date from that time but also from the following phytogeographical evidence. In the four great temperate land masses of the Southern Hemisphere (Australia, New Zealand, temperate South America, and South Africa) the endemic genera and the very characteristic families, which undoubtedly represent the most ancient element in the vegetation, are composed, in very large majority, of woody plants,[1] as shown in Table II. Most of the indigenous herbs, however, belong to genera, and often

TABLE II

	Total	Herbs	Percentage Herbs
Australia—			
Species of non-endemic genera	2,301	1,251	54
Species of endemic genera	4,024	677	17
New Zealand—			
Species of non-endemic genera	711	460	64
Species of endemic genera	315	109	34
Patagonia—			
Species of non-endemic genera	920	720	76
Species of endemic genera	667	320	48
South Africa—			
Species of non-endemic genera	3,298	1,929	55
Species of endemic genera	4,686	1,390	29

[1] "Endemic" genera are those which are either strictly confined to the region in question or have only a very few species outside of it.

even to species, which are identical with those in northern latitudes, and are therefore in all probability recent immigrants.

The slight degree of endemism among these southern herbs is the more significant when we remember that such plants, from the brevity of their life-cycle, are likely to change much more rapidly than woody forms and hence would develop into endemic types in a shorter time. The notable uniformity of flora, too, over most of the globe during the Cretaceous, as shown by fossil evidence, renders it very improbable that, if herbs had been a dominant feature of the vegetation of that period, they would now be so uniformly absent from the ancient portion of the southern floras. The union of southern Africa with Eurasia in the Miocene and the union of South America with North America in the Pliocene evidently mark the periods of the herbaceous invasion of these two continents. Australia has apparently been sufficiently isolated from Asia since the Cretaceous so that land animals have been unable to enter it. The northern members of its flora, however, have doubtless effected their entrance by a migration, in comparatively recent times, over the Himalayas and along the East Indies. Had herbs been numerous in the latter part of the Mesozoic, when the connection of Australia with Eurasia was evidently much more intimate, the present vegetation of the island continent would certainly contain a much higher percentage of such plants. That the very early Tertiary flora of Australia included few if any herbs is made even more certain by the fact that although in all probability northern Australia and New Zealand were connected at that period by a land bridge, none of the northern herbs which entered Australia from the East Indies are now found in New Zealand; and we are obliged to infer that they had not then arrived in Australia.

The period at which herbs became an important feature of the north temperate vegetation and began to spread thence southward is therefore pretty clearly set at somewhere in the early Tertiary.[1] That herbaceous dicotyledons existed in the latter part of the Mesozoic, however, especially in mountainous regions, is most

[1] Recently discovered evidences of glaciation just at the close of the Cretaceous are of importance in this connection.

probable, but it seems equally probable that their great development did not take place until after the close of that epoch.

Such, in brief, is the evidence for believing that herbaceous angiosperms are comparatively recent in origin; that they have been developed most abundantly in the north temperate zone, mainly since the Cretaceous, as an adaptation to the advent of a winter season; and that they have spread thence over most of the other regions of the earth. Let us now see what conclusions of importance to geologists and climatologists may be drawn from a study of the origin and dispersal of these herbs.

In the first place the distribution of herbs in the north temperate zone provides us with evidence as to the climate of this area at the time of the last Glacial epoch.

As to what was the composition of the northern flora before the advance of the ice sheet we cannot be very certain, but there is reason to believe that a much higher proportion of woody plants flourished there than at present, indicating the existence of a climate devoid of extreme cold. Such an inference is based on the composition of that considerable body of related species and genera which are common to eastern North America and eastern Asia and which are nearly or quite absent elsewhere. Gray[1] was the first to suggest that these plants were the remnant of the widespread pre-Glacial boreal flora which had been forced south by the advance of the ice and had been exterminated almost everywhere save in these two regions, which were the only ones in warmer latitudes into which it had been able to escape. If this hypothesis is correct, as subsequent investigations seem to indicate, and if this group of plants is truly a fair sample of the pre-Glacial flora of the north, a study of its composition is of considerable interest. Gray has published a list of the genera and species of eastern North America which are absent in Europe but which are represented by identical or closely related forms in eastern Asia. This list comprises 142 genera of dicotyledons, of which 70 are woody or predominantly so, and 240 species, of which 128 are woody. The flora is thus just about equally divided between herbs and woody plants. From the greater hardiness of most herbs it is probable that they have

[1] A. Gray, *Scientific Papers* (1879).

suffered less by extinction than have the woody plants, and consequently that even more than 50 per cent of the original boreal dicotyledonous flora consisted of woody forms. This analysis would seem to indicate that previous to the glacial invasion the climate of the north temperate zone had, for a long time at least, been devoid of extremes of cold. We know from fossil evidence that in several regions during the Tertiary temperate genera, such as oaks and poplars, grew side by side with such tropical or subtropical types as the palm and fig; and it therefore seems reasonable to infer that the climate was a very equable one and devoid of extremes of heat as well as of cold.

More conclusive evidence, however, is at hand as to climatic conditions near the ice front during the actual period of the glacial invasion. The vegetation of the whole temperate zone at that time of course lost heavily by extermination, but this extermination must have been much more pronounced among woody plants than among herbs, owing to the greater ability of the latter to withstand cold and other adverse conditions. It is a significant fact that the present flora of Europe north of the Alps is decidedly more impoverished than is that of corresponding temperate North America, and that although in the latter region approximately 25 per cent of the dicotyledons are woody, in the former only from 10 to 15 per cent are so. This paucity of indigenous trees and shrubs in northern Europe is especially noteworthy since experiment has shown that many species of delicate and warmth-loving trees and shrubs will grow in England, France, and Germany which cannot stand the droughts and winters of the northern United States. In America, too, there are representatives of a considerable number of woody families which are now absent in northern Europe, many of which occur there as fossils. These facts are evidently to be explained by the much greater adversities suffered by the European flora during the Glacial epoch. In North America, especially in its eastern portion, the vegetation could easily migrate southward at the advance of the ice and return northward at its retreat. In northern Europe, on the other hand, the southward escape of the vegetation was blocked, and it was crowded against the Alps, the Pyrenees, and the Mediterranean, thus suffering

heavily by extinction. These same natural barriers have also prevented any considerable northward migration since the retreat of the ice. The vegetation of northern Europe today seems, therefore, to be descended directly from that remnant which was able to survive on the unglaciated portions of France, Germany, and England. We have already noted the fact that the percentage of woody plants in the dicotyledonous flora of northern Europe is amazingly low, being only from 10 to 15 per cent of the whole. This is really a percentage typical of alpine or northern regions. Indeed, the proportion of herbs in Switzerland today is but little higher than that in the adjacent lowlands. The alpine character of the northern European flora is further emphasized by the strong resemblance which it bears to that of the flora of the Rocky Mountains, for the two floras are composed of almost exactly the same families and include a host of identical genera and even a large number of identical species. In fact, the flora of the Rockies presents a much closer resemblance to that of Europe than does the flora of the eastern part of the continent.

If the flora of northern Europe is indeed typically representative of that which flourished near the ice front during glacial times, the proportion of woody forms within it affords us a valuable index as to climatic conditions during the height of the ice age. The facts seem to indicate that when the ice sheeet had reached its greatest extent the country in its immediate front was neither a barren arctic tundra, as has sometimes been supposed, nor covered with a luxuriant temperate vegetation; but that the climate in general resembled that of the lower portions of the Alps or the Rockies today, being cold enough in winter to kill off all but the hardiest trees and shrubs but not sufficiently cold to reduce the whole vegetation to the few perennial herbs and stunted shrubs which are characteristic of arctic regions today. Of course this evidence is of value only as indicating the climate during the *coldest* period of the ice invasion, just as the percentage of herbs in a flora is indicative of the minimum winter temperature of the region. As to the climate of the presumably warmer interglacial periods it tells us nothing.

This conception of the dicotyledonous herb as a comparatively recent type of plant which has been developed extensively in the north temperate zone during the Tertiary and has spread thence far and wide over the globe also leads us to some interesting geographical conclusions. Those regions, such as southern Asia and Mexico, which have been invaded, at least in their more temperate portions, by a flood of northern herbs, have in all probability had a nearly or quite continuous land connection with northern Eurasia or North America during Tertiary time. Other regions, such as Australia, New Zealand, South America, and South Africa, where there is a markedly smaller percentage of herbs in the vegetation, have apparently had a much less intimate connection with the great land mass of the north temperate zone during the Tertiary. This is particularly true of certain oceanic islands, notably Hawaii, the Polynesian group, Juan Fernandez, the Canaries, St. Helena, Socotra, Mauritius, and the Seychelles, on which the percentage of herbs is much smaller than on the adjacent mainland. In these insular floras practically all the herbs belong to species which also occur on the near-by continent; and almost the entire body of the endemic genera, presumably the most ancient portion of the flora, is composed of woody plants, as is shown in Table III.

TABLE III

	Total	Herbs	Percentage Herbs
Hawaii—			
Species of non-endemic genera	339	117	34
Species of endemic genera	243	21	8.6
Fiji—			
Species of non-endemic genera	543	81	15
Species of endemic genera	20	0	0
Juan Fernandez—			
Species of non-endemic genera	57	27	47
Species of endemic genera	17	0	0
St. Helena—			
Species of non-endemic genera	34	15	44
Species of endemic genera	7	0	0
Mauritius and the Seychelles—			
Species of non-endemic genera	524	191	36
Species of endemic genera	63	3	4.7

This paucity of herbs, particularly of annuals, on oceanic islands was noted long ago by Darwin, Hooker, and others, and was ex-

plained (by Darwin[1]) as due to the fact that the lessened competition between the members of an insular flora allowed many plants to grow there into shrubs and trees which on the mainland could never succeed in attaining more than a herbaceous stature. From what we have seen as to the evolution of herbs this explanation appears inadequate, and it implies that, in many cases, herbs must have lost their great advantage of being able to pass from seed to seed in a single season. Insular floras appear rather to be relics of the ancient Tertiary vegetation which once flourished on the adjacent mainlands but which has been more or less superseded there by an influx of new plants, most of them herbs. The many striking differences in flora between Juan Fernandez and adjacent Chile; the Canaries and adjacent Morocco; St. Helena and adjacent South Africa; and Socotra and adjacent Somaliland seem to point to this conclusion. Strongly in favor of such a view are also the very numerous floral resemblances exhibited between distant islands or between islands and distant continents. The occurrences of shrubby species of Plantago only in Hawaii, Juan Fernandez, and St. Helena; of phyllodineous Acacias only in the Mascarene region, Hawaii, and Australia; of related shrubby Compositae in Hawaii, Tahiti, the Galapagos, and Juan Fernandez, and of numerous related species in the Canaries, South Africa, and Socotra constitute a few of many instances of such distribution. The very large proportion of woody plants compared to herbaceous plants in these insular floras and the high degree of endemism of the former as opposed to the latter strongly favor the theory that these particular islands do indeed support a very ancient type of organic life, and that since the opening of the Tertiary at least they have not been intimately connected with any large continental area.

Other isolated islands or archipelagoes, such as Bermuda and the Azores, possess as high a percentage of herbaceous plants in their vegetation as do the adjacent continental areas, and they likewise have a very small endemic element. We are forced to conclude, on evidence both from the flora and from the composition of the vegetation, that such islands have appeared, or at least have received their plant life, in comparatively recent times.

[1] C. Darwin, *Origin of Species*, 6th ed., p. 413.

Aside from its bearing on the antiquity of oceanic islands and other isolated regions, the present theory as to the origin and dispersal of herbs is of importance as indicating the conditions under which occurred that momentous botanical event of the Tertiary—the invasion of the Southern Hemisphere by a flood of northern plants. Phytogeography indicates that this invasion took place along three main routes: over the Andes into Patagonia and thence over an extensive antarctic continent or archipelago into New Zealand and southeastern Australia; over the central African highlands into South Africa and Madagascar; and over the Himalayas and along the East Indies into Australia. These invasions seem to have ceased entirely at the present time, for there are in most cases wide gaps of hundred or even thousands of miles between the northern and the southern ranges of a genus or a species. That the immigration into Africa and Asia ceased some time ago is indicated by the fact that in these continents there are now almost no species identical with northern ones, although the genera are still the same. The highway over the Andes, however, seems to have closed much more recently, for there are a large number of species which traversed it that are still identical with their boreal types. The significant facts in this connection are that, with the exception of a very few genera which are poor in species, this invasion by the "Scandinavian" flora was an invasion of *herbs*,[1] and that herbs are peculiarly adapted to temperate climates. This immediately suggests the conclusion either that these three mountain highways were considerably more elevated during the Tertiary, with a consequent increase in the extent of temperate areas (a

[1] The presence of such temperate types as Betula, Populus, and Quercus in the Cretaceous of Patagonia, as described by F. Kurtz ("Contribuciones a la palaeophytologia Argentina—Sobre la existencia de una Dakota flora en la Patagonia austro-occidental," *Revista museo la plata*, X [1899], 1902, pp. 43–60), and in the Cretaceous and even Tertiary of Australia and New Zealand by Baron C. von Ettinghausen ("Tertiary Flora of Australia," *Gov. Surv. N.S.W. 1888*, p. 82; "Contributions to the Knowledge of the Fossil Flora of New Zealand," *Trans. New Zealand Inst.*, XXIII [1890], 237) (the latter's identifications are often open to grave question, however) has probably nothing to do with the Tertiary invasion of herbs but rather bears witness to the remarkable uniformity and intermingling of the Cretaceous flora. Had these ancient temperate plants been accompanied by anything like the throng of herbs which now surrounds them in the north, it is highly improbable that herbaceous forms would now be so scarce in the south temperate zone. Because of the brevity of their life-cycle, herbs are likely to change very rapidly, and the slight degree in which these "northern" herbs in the antipodes have become altered also bears witness to their comparatively recent arrival.

hypothesis which the Tertiary origin of the Andes and Himalayas may perhaps be regarded as supporting); or that for some other reason, possibly an increase in glaciation, the area of regions enjoying a cool climate in the tropics has at various times been considerably augmented.

In this connection it is of interest to note the evidence which is accumulating as to the occurrence of more or less widespread glaciation both at the transition from Cretaceous to Tertiary and again during the Miocene. It is possible that the more ancient herbaceous invasions were coincident with these earlier glacial periods and that the later migration of such plants into the antipodes was contemporaneous with the extensive glaciation which marked the latter part of the Tertiary.

The theory put forward by Hackel[1] and supported by Scharff[2] asserts that the boreal plants in antipodean South America are extremely ancient types which have preserved their specific identity for an exceedingly long time; that they migrated over a land bridge long since destroyed; and that instead of having come south along the Andes they are at present going north. Such a theory is certainly not tenable if herbs are proven to be of recent origin. The increasing frequency of northern herbs as one goes south along the Andes, the fact on which Hackel's hypothesis is based, is well explained by a decrease in temperate areas within the tropics as by a migration from the south. It is also very difficult to imagine so many forms maintaining or almost maintaining their specific identity for such a very long period of time. The present theory of the origin of herbs certainly supports the view that they have entered the antipodes rather recently over land bridges which have not had time to become fundamentally altered.

We have already spoken of the antarctic continent as a highway between South America and Australasia; and the existence in comparatively recent times of such an extended Antarctica, able to support a large fauna and flora, has been postulated by almost all students of the distribution of antipodean plants and animals.

[1] E. Hackel, "Über die Beziehungen der Flora der Magellansländer zu jener d. nördl. Europa und Amerika," *Mitteil. naturw. Verein Steiermark* (Botan. Sekt.) (1905), pp. 110–15.

[2] R. F. Scharff, *Distribution and Origin of Life in America* (1912), p. 418.

That during its greatest extent, however, it was probably not a continuous land bridge but more in the nature of a great archipelago, presenting a barrier to most animal invaders but much more easily crossed by plants, seems to be generally admitted. It is worth our while to analyze the ancient "endemic" flora of this antarctic continent and archipelago and to find, if we may, what the climatic conditions were under which it flourished. It seems to be a reasonably safe conclusion that all genera commonly designated as "antarctic," from their confinement to the temperate regions of the Southern Hemisphere, were once inhabitants of Antarctica. The writers have compiled a list of eighty-eight dicotyledonous genera which have representatives in at least two of the three main antarctic regions—New Zealand, Australia, and temperate South America—and which possess but very few species outside these regions. This list may well be regarded as representative of the flora of Antarctica (except for the northern invaders) before the advent of the cold period which drove the phanerogamic vegetation northward into South America and Australasia. Of these eighty-eight genera only thirty-four, or 38 per cent, are typically herbaceous. These few herbs are obviously unrelated to northern ones and seem clearly to have had an independent origin in the antipodes in response to the refrigeration of the climate. That they are not as ancient as the northern herbs appears to be indicated by the fact that the annual type has not yet been evolved among them. If the northern invasion which crossed Antarctica did not enter it until the Pliocene (as seems probable), no very extreme refrigeration could have taken place until the latter part of that period, at any rate. The very high percentage of woody plants in the original flora, however, seems to testify strongly to the existence in Antarctica at no very ancient date of a climate devoid of extreme cold. The refrigeration of this climate and the evolution of its herbaceous vegetation, instead of being the slow and gradual processes that they were in the Northern Hemisphere, seem to have been much more rapid. The consequently scanty number of herbs and the lack of time or space for very vigorous competition among them doubtless explain why the antarctic herbs are not as widespread and aggressive as their northern congeners.

Such are the more important geological and climatological inferences to which we are led by a recognition of the recent and predominantly boreal origin of herbs, and of the factors which have determined their origin. Much more complete data must be sedulously gathered, particularly by the phytogeographer and the botanical phylogenist, before the full application of this theory to the problems of geology can be definitely determined. That it is possible, however, to draw conclusions of importance to geology and climatology, not alone from the past and present distribution of species, but also from a study of the evolution of the growth of habits of plants, will be readily admitted.[1]

SUMMARY

1. The earliest angiosperms were woody plants, and herbs have been derived from them by reduction. This is indicated by the facts that: (1) the earliest angiosperm fossil remains are almost entirely of woody plants; (2) the uninterrupted and active cambial ring of woody dicotyledons, which lays down secondary wood provided with well-developed medullary rays and wood parenchyma, is the primitive type and that the much reduced secondary tissues in herbs are clearly derived from this; (3) the ancestors of the angiosperms and the great majority of the primitive members of the phylum are woody plants; and that (4) in all families and genera which contain both woody plants and herbs the latter are more primitive in their constitution.

2. The differentiation and refrigeration of the climate of the temperate zones since the beginning of the Tertiary, and the consequent appearance of a cold winter season, have been the chief factors which have caused the evolution of herbs. Herbs are able to survive adverse conditions of temperature or moisture as underground roots and stems or in the form of seeds, and can thus thrive in regions where plants with perennial aerial stems would perish. In general, the lower the winter temperature of a region the fewer the woody plants in its flora.

3. At least half of the pre-Glacial vegetation of the north temperate zone seems to have been composed of woody plants,

[1] A more complete discussion of the present problem with the presentation of a much larger body of data will be found in a paper by the writers, "The Origin and Dispersal of Herbaceous Angiosperms," *Annals of Botany*, XXVIII (1914), 547–99.

indicating the occurrence of a rather mild winter. The advent of the Glacial period resulted in very great extermination of the flora, but herbs suffered much less than did woody plants.

4. In America the vegetation was able to return northward after the retreat of the ice and thus shows a considerable proportion of woody species today. In northern Europe, however, natural barriers have to a great extent prevented this return, and the present flora of that region seems to be descended from the remnant of the pre-Glacial vegetation which survived on the unglaciated areas. The fact that the proportion of herbs in the present north European flora is like that in northern or low alpine regions today provides us with a clue as to climatic conditions during the height of the glacial invasion.

5. The land mass of the north temperate zone has been the seat of origin of a very large part, if not of the majority, of herbaceous genera. The opportunities for wide dispersal and keen competition here have resulted in the production of a very hardy and aggressive herbaceous flora which has spread widely into the tropics and the Southern Hemisphere.

6. This invasion of herbs from the north has taken place for the most part during the Tertiary, and the percentage of herbs in the floras of the various regions in the tropics and the Southern Hemisphere indicates roughly the intimacy of the connection between these regions and the land mass of the north temperate zone during Tertiary time. The floras of certain oceanic islands, for example, seem from the extreme paucity of herbs to be very ancient in type, implying a very long period of isolation; whereas the floras of other islands, overwhelmingly herbaceous, indicate a recent origin for these islands or at least for their plant populations.

7. Boreal herbs in the antipodes are usually separated from their northern congeners by considerable distances, indicating that migration has ceased and that the wide area of rather temperate climate within the tropics, under which the invasion took place, has recently become much smaller.

8. A reconstruction and analysis of the ancient flora of the antarctic continent indicate that up to comparatively recent times its climate was very mild and that refrigeration took place more rapidly than in the north temperate zone.

POST-CRETACEOUS HISTORY OF THE MOUNTAINS OF CENTRAL WESTERN WYOMING

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PART III

Quaternary cycles of stream erosion.—Although in the higher mountains stream erosion was superseded at intervals by glaciation, over most of the district the Quaternary period was a time of continuous action by running water and its accessory agencies. The erosion history was accented, however, by several changes which rejuvenated the streams from time to time, and thus prevented the completion of the successive erosion cycles. Conceivable causes of such interruptions are: (*a*) a series of general uplifts; (*b*) the readjustment of stream grades on account of the removal of obstructions in the channels of the master rivers, or because of piracy in their lower courses; and (*c*) notable changes of climate.

Of these causative changes, those which were climatic should have produced more or less harmonious results simultaneously over the entire district. The same should have been true of the gentle elevatory movements, if they were uniform over wide areas. The two factors should differ in their results in that climatic changes would doubtless make themselves felt all over the district at the same time, and streams would then respond either by trenching their floodplains or by aggrading them. The changes would apparently take place simultaneously throughout the entire graded profiles of all the streams. The elevatory movements, however, if equal everywhere, would cause rejuvenation first in the lower courses of the streams, or, in this instance, outside the district. The phenomena due to elevation or depression would then migrate up the individual streams; in such cases the new cycle features of the lower courses would be distinctly older than those of the upper courses at any given time. If the surface of the district, instead of

being uniformly raised (relatively), was warped, then the readjustments of stream action and their results must have been more complex. Rejuvenation due to the removal of falls and rapids or to stream-piracy would be likely to cause changes in one valley system which would have no counterpart in the adjacent drainage basins.

Climatic change may well explain some of the minor rejuvenations, but apparently is incompetent to account for the cutting of valleys hundreds or even thousands of feet deep. Inasmuch as it is not only possible, but fairly probable, that several or even all of these changes have been at work, it is clear that the problem to be dealt with is by no means simple; it should be approached in a spirit of caution and without an expectation of solving it completely except after more detailed field studies. It is an observed fact that the topographic features in different parts of the district here considered, and even in entirely different drainage basins, are so similar in aspect, in order, and in relations as to suggest that each has passed through a similar history, owing to the operation of some widespread influence. These facts suggest either climatic changes or regional uplifts. There is no question as to the competence of uplifts to induce dissection to proportionate depths, but climatic changes competent to bring about the incision of valleys to depths of hundreds or even several thousands of feet, as is indicated by the canyons in the ranges of Wyoming, would have to be extraordinary in magnitude. For that reason the hypothesis of changes of level seems to me the more favored one, although it should not be inferred that the other influences are excluded from participation in a minor degree.

The work of interpreting the erosion history of such a district involves largely the recognizing and discriminating of surfaces produced during successive cycles and correlating these in chronological order. Although it is often possible to recognize mature and even young topographic forms of preceding cycles of erosion, the most readily recognized features are the broad graded surfaces or flood plains which normally become wider and wider as the cycle proceeds. In Wyoming, however, a careful discrimination must be made between remnants of old river-graded surfaces, and those

other flats and terraces which are normally developed by the stripping of weak strata from harder beds in horizontal position—a process which is peculiarly effective in dry wind-swept plains. This distinction must be made particularly in the broad basins such as the Wind River and Green River basins, where the Tertiary strata have been but little disturbed. It is necessary also to bear in mind that the graded surfaces, which constitute the immediate goal of all streams, are not level, either transverse to the main stream or parallel to it. Therefore remnants of the same graded surface generally do not have the same elevation in different parts of the district, even though there has been no subsequent warping; but such remnants should bear approximately the same general relation to each other as corresponding parts of the modern floodplains bear to each other.

The studies of Westgate and Branson, already noticed, show that in the Lander district the topographic features reveal a definite sequence of cycles which may eventually be extended to a much larger region. The existing topography suggests a declining series of phases, in which the earliest cycles proceeded to old age, the intermediate ones to maturity, and the later ones only to youth. This may well be more apparent than real, however, for if any of the early stages had gone no farther than youth, the evidence of their existence would have been destroyed in those later cycles which went farther toward completion. As I am still somewhat doubtful about the correlation of the features I have seen with those described by Westgate and Branson, I present in Table II an independent scheme of glacial stages and erosion cycles, not as a final list, but necessarily as a provisional one.

Below the elevated peneplain, which has already been described as probably of Pliocene age, I find evidence suggesting at least four cycles of erosion, one of which appears to be older than the oldest glacial drift, while the rest are more or less associated with glacial phenomena. The latest cycle, which is probably complex in detail, is represented by the inner valleys with existing floodplains and their immediate terraces. It was closely associated with the last glacial stage. The phenomena of the various cycles may now be discussed more fully.

Fremont cycle: The oldest cycle recognized, of which the visible result was the "Wind River peneplain," may for the sake of uniformity be called the Fremont cycle, from the fact that the principal remnants of the Wind River plateau (Figs. 12–15) may be overlooked from Fremont Peak. In the Wind River Range the remnants of this surface attain elevations of 11,300 to over 12,000 feet, and in adjacent regions to the west other remnants supposed to be of the same age are between 10,500 and 11,000 feet above sea-level. Its features have been sufficiently described in *Jour. Geol.*, XXIII (1915), 193–94.

TABLE II

Epoch	Erosion Cycles (Largely Interglacial)	Glacial Stages	Stages according to Westgate and Branson
Recent	Post-glacial	....	Late glacial and terraces
Pleistocene	....	Pinedale	
	Lenore	....	
	....	Bull Lake	Early glacial
	Circle	....	Plain No. 1
	....	Buffalo	....
	Black Rock	....	Plain No. 2
	Union Pass	....	Plains Nos. 3 and 4
Pliocene	Fremont	....	Summit peneplain

Union Pass cycle: The next stage in the denudation is typically represented by a post-mature surface cut upon the resistant Archean and Paleozoic rocks near the west end of the Wind River Range notably below the summit peneplain. It is evidenced by the broad, shallow, and well-graded valleys with elevations of from 9,000 to 9,500 feet at Union Pass and west of Sheep Mountain. East of Kendall, Little Flat-Top Mountain, of about the same elevation, is a beveled remnant of highly inclined Cretaceous rocks. Along the northeast slope of the Wind River Range, aged topographic forms high up on the mountain sides probably represent

this cycle, although the correlation is not assured. At the southeast end of the range, Westgate and Branson[1] distinguish two erosion cycles beneath the summit peneplain and higher than the cycle represented by Table Mountain near Lander, which I have tentatively correlated with the Blackrock cycle, farther east; these do not appear to be as distinct in the region I have studied, and were not recognized.

Still farther westward, there are, at a comparable elevation, broad, open, and well-graded valley heads both east and west of Twogwotee Pass, but always deeply intrenched farther downstream. Aged forms of similar aspect and relations have been observed between the valleys of Jack Pine and Conant creeks in the northwest part of the Teton Range. This gently sloping surface lies decidedly higher than the oldest deposits of glacial drift, but is a thousand feet or more below the tabular remnant interpreted as possibly part of the Goat Flat peneplain. In fact, the whole western slope of the Teton Range in the vicinity of the north fork of Pierre's River is a gently inclined plain rising from the general plateau level of about 6,000 feet to the bases of the more rugged peaks at 9,000–9,500 feet. Although it has been incised by deep canyons, there are still broad smooth remnants on the divides. The fact that the canyons are much deeper near their heads than farther west suggests that the surface has been tilted by being raised on the east. Although it is underlain by rocks of various types and structures, the plain is but little modified by structure.

At the west end of the Gros Ventre Range a few miles northeast of Jackson, there are rather large remnants of an ancient gentle valley slope cut across folded Paleozoic rocks. These remnants stand nearly 2,000 feet above the channels of the present streams.

No doubt the surface produced in the Union Pass cycle was broadly planed on the weak Mesozoic and Tertiary clays, although only late mature topography developed on the harder rocks near the headwaters of the streams, and some remnants of the summit peneplain were left untouched. In this cycle the latter was

[1] L. G. Westgate and E. B. Branson, "Cenozoic History of the Wind River Mountains," *Jour. Geol.*, XXI (1913), 142–59.

probably reduced very nearly to its present remnants, although its relative elevation has since been greatly increased.

Black Rock cycle: The next cycle has been named after Black Rock Meadows, near the head of one branch of Snake River, which at that point has excavated a flat-bottomed valley with gentle slopes chiefly on Mesozoic strata. Farther downstream, the flat is trenched and appears as a series of flat-topped spurs. These features lie several hundred feet lower than the valleys representing the Union Pass cycle in the vicinity. More or less generally throughout the district, the Black Rock valleys and remnants are 500–800 feet above the present stream. Along the Gros Ventre River, the cycle is probably represented by the long flat-topped spurs from one to eight miles below the mouth of Fish Creek. The surfaces of these spurs trunkate Cretaceous beds which dip from 15° to 25°, and if the intervening gulches were filled there would be a continuous graded valley floor several miles wide. On one of the remnants a thin capping of water-worn gravel is still preserved. Corresponding features may be found in the tabular areas and flat gravelly spurs which obstruct the Snake River valley below Jackson Hole at elevations of 500–600 feet above the river, and the similar forms in the lower Snake River valley near Irwin, Idaho. These forms are apparently remnants of a wider, flat-bottomed valley much shallower than the present canyon of Snake River. In the Teton Range the wide shallow valleys of similar nature, from 500 to 1,000 feet above the bottoms of the modern canyons, suggest the same cycle.

Returning eastward to the north side of the Wind River basin, we find around Black Mountain (see Fig. 24), and westward along the mountains to DuNoir River, a high plain, distinctly lower than the Union Pass cycle, and now much dissected. Most of these remnants are veneered with weathered stream-gravel which plainly indicates the origin of the plain. At the mouth of each ravine in the mountain side, an old alluvial fan is still to be seen although nearly all such fans have been dissected. In some places in this district there is evidence that there are really two plains separated by a terrace 50–100 feet high, but in general there is not sufficient basis available for resolving the Black Rock cycle into two. In the

center of the basin the tributaries have largely destroyed the plain, leaving only small mesas or isolated buttes such as Crowheart Butte.

Around Lander and the southeast end of the Wind River Range, there are mesas and shelves along the foothills cut across inclined strata, at elevations of 600–700 feet above the present streams (Figs. 30, 31, 32). These are considered by Westgate as remnants of his plain No. 2, and they also have the characteristics and relations appropriate to the Black Rock cycle as defined in preceding

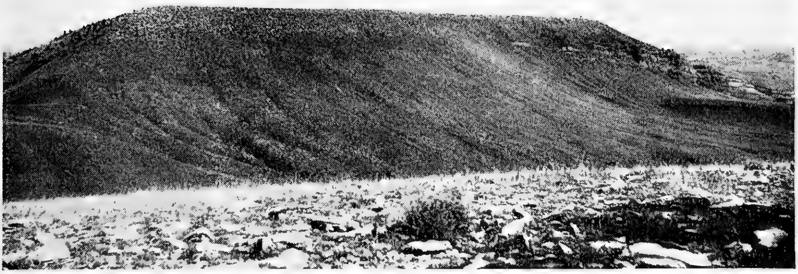

FIG. 30.—Table Mountain, near Lander. A mesa cut on tilted strata and capped with gravel. (Photograph by D. D. Condit, U.S. Geological Survey.)

pages. The layer of coarse gravel 50–250 feet thick in places, forming the top of each of these remnants, has interested geologists since the time of the Hayden Survey. On account of the fact that a few of the bowlders are as much as 25 feet in diameter, and many of them 5–10 feet thick, it has been suggested that the deposit must be glacial. Westgate, however, cites the distinct although crude stratification of the material as evidence that it is stream-laid. Even larger bowlders have been carried out from the canyons of the Sierra Nevada by torrents.[1] The hypothesis that these

[1] A. C. Trowbridge, *Jour. Geol.*, XIX (1911), 706–47.

patches of gravel near Lander are remnants of torrential fans agrees with the evidence so far as I have seen it.

In the Owl Creek Range also features indicating the Black Rock cycle have been observed although but sparingly. South of Anchor certain shoulders 600–800 feet above the valley bottoms are strewn with coarse, deeply weathered gravel and bowlders, which were once water-worn. There is considerable resemblance between conditions here and near Lander. In the upper course of Green River

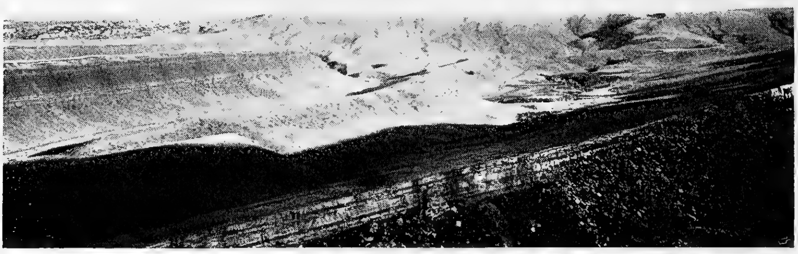

FIG. 31.—Tabular divide about twenty miles south of Lander. Probably represents the Black Rock cycle. Hogbacks of Permo-Mesozoic strata in the foreground. (Photograph by D. D. Condit, U.S. Geological Survey.)

similar high buttes and mesas, such as Flat Top Mountain near Kendall, suggest the same cycle.

In the type locality, the Buffalo old drift seems to lie always upon the remnants of Black Rock valley floors, and in the Teton Range the relations are similar. From these facts I infer that these shallow older valleys were completed about the time of the first well-recognized stage of glaciation in western Wyoming. It is a persistent and distinctive characteristic of the Black Rock surface that it is floored with thick and very coarse gravel deposits. None of the earlier or later partial plains of the region carry more than a thin veneer of such material. It has been suggested that the

coarse, stratified gravel deposits represent outwash from the ancient glaciers rather than torrential fan deposits under the influence of an epoch of desiccation. With reference to deposits now so deeply weathered, I know of no criteria that could be used to decide between these two modes of deposition. Whatever the origin, there is a suggestive resemblance between these thick coarse gravel deposits on remnants of the Black Rock plain and similar deposits in other ranges of the Rocky Mountains, such as the conglomerate

Fig. 32.—High rock-cut terrace southwest of Lander. The Carboniferous strata have been planed off at the level of Table Mountain (see Fig. 30).

of Bald Ridge on the east side of the Bighorn Range,[1] the Bishop conglomerate of southwestern Wyoming,[2] and other well-known examples.

Between the Black Rock cycle and the next one clearly discriminated there may well have been one or more cycles now represented by terraces visible here and there in the Wind River and Green River bad lands. These are, however, but little known and

[1] *U.S. Geol. Survey, Cloud Peak–Fort McKinney Folio*, No. 142.

[2] J. L. Rich, *Jour. Geol.*, XVIII (1910).

appear not to have left notable marks in any but the softest strata. For the present, therefore, they must be neglected.

Circle cycle: Benches cut across folded rocks along the sides of the mountain valleys and much wider terraces and mesas cut in the Tertiary basin are relics of a surface carved out of the Black Rock peneplain. Near the mouth of Fish Creek, the principal northern tributary of the Gros Ventre River, a broad terrace (Fig. 33) 100–200 feet above the stream is very conspicuous.[1]

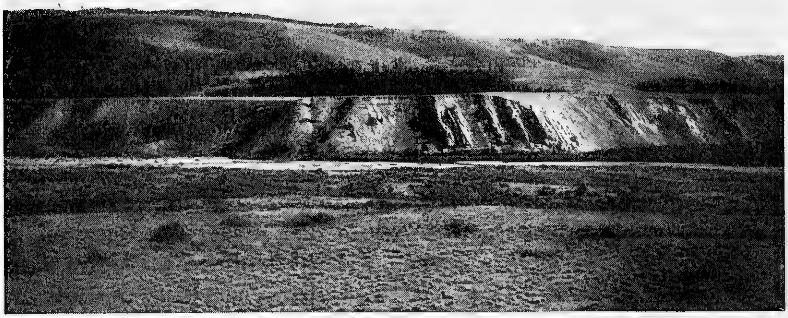

FIG. 33.—Rock-cut terrace on Fish Creek, correlated with the benches at Circle. The Cretaceous strata are nearly vertical.

To the east, in the upper part of the Wind River basin, there is a prominent terrace cut on inclined strata and again standing about 150 feet above the stream (Figs. 34 and 35). A fine remnant of it stands out just east of Circle post-office on Wind River. In each case the surface is strewn with coarse, river-deposited gravel, but the layer is only a few feet in depth. As the terrace blends into the earlier moraines of the Dinwoody and Torrey glaciers, the gravels may well consist largely of the outwash deposits of the Bull Lake stage. Again, in the vicinity of Bull Lake and Fort Washakie, a nearly smooth upland surface has been planed across tilted

[1] *U.S. Geol. Survey*, Mount Leidy, Wyoming, topographic sheet.

strata at about the same height above the creek. Other remnants (Fig. 36) may be seen along the slope of the Wind River south-

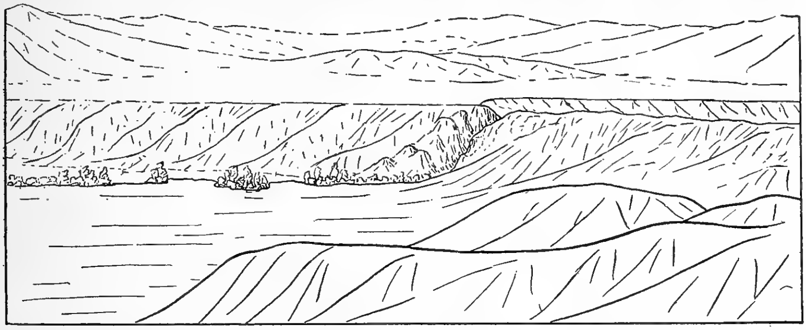

FIG. 34.—The Circle terrace near Duncan's Gorge on the North Fork of Wind River. (Drawn from a photograph.)

FIG. 35.—Rock-cut terrace of the Circle plain on Wind River at the mouth of Red Creek. Triassic (?) beds dip to the right.

west and south of Lander. This is the plain No. 1 of Westgate and Branson.[1]

[1] *Jour. Geol.*, XXI (1913), 156.

Near Anchor, on the north side of the Owl Creek Range, the most conspicuous tabular hills stand about 150 feet above the present valley bottoms (Fig. 37).

In Jackson Hole the low hills scattered about the southern end of the basin are not flat-topped, but their surfaces are only gently undulating and appear to represent late mature topography of an earlier cycle. Corresponding to the tops of these hills are wide flat-bottomed valley heads in the adjacent foothills, and all about

FIG. 36.—Mesas and terraces of the Circle cycle cut across inclined Mesozoic strata in the Shoshone Indian Reservation. The higher surface at the right may represent the Black Rock cycle.

150–400 feet above the bottom of Jackson Hole. These and similar features along the west slope of the Teton Range may correspond in age and represent the Circle cycle, although the correlation at such a distance is far from assured. Northeast of the new town of Tetonia, at the north end of the Teton basin in Idaho, there are low, nearly flat-topped mesas which have been dissected to a depth of 100–200 feet by the streams which excavated the bottom of the Teton basin. These features again are of the same order of magnitude, and have the same general relations as those typical of the Circle cycle. The Circle floodplains were once two

miles wide at the type locality, but were much broader near Fort Washakie and Lander. In the somewhat more remote headwaters of the Gros Ventre valley, the terraces are only a fraction of a mile in breadth. They are distinctly older than the latest (Pinedale) stage of glaciation. The older (Bull Lake) moraines, however, rest upon the Circle terraces and blend into them in such a way as to indicate that the terraces were parts of the floodplains in the earlier glacial stage.

FIG. 37.—Circle terraces along North Fork of Owl Creek. Bedded Tertiary volcanics in the background.

Lenore cycle: Following the last distinct rejuvenation, the inner trenches of Wind River and its tributaries were excavated in the Circle floodplains (Figs. 39, 40). Although these valleys are relatively narrow in their upper reaches, they are well graded, and the floodplains are broad enough locally to accommodate ranches, as at Lenore and Circle. They expand considerably downstream.

Farther west in the Gros Ventre valley there is a similar inner valley cut in the well-defined terrace of the Circle cycle. Fish Creek has intrenched the Circle plain to a depth of 100–150 feet, and then by meandering has developed a floodplain about 1,000 feet wide (Fig. 41). On the whole, the valleys are considerably narrower here than in the Wind River basin, but the rocks

are somewhat more resistant and the region is closer to the headwaters of the streams. Westward this inner valley merges into the floor of Jackson Hole. Since the latter is now coated with recent alluvial deposits of unknown depth, it may prove to be somewhat unlike the Wind River basin in origin. If the filling is deep it may signify local warping or faulting, and the appropriate disturbance of stream gradients; but if it is thin, as I think more probable, it may be ascribed entirely to the glacial choking of Snake River.

Inner trenches of varying width, but generally similar to those farther east, are observable along Fall or "Hoback" River, down the Grand Canyon of Snake River to its lower valley near Irwin, and also along its southern tributary, Salt River, as far up as Star valley, in Wyoming and Idaho.

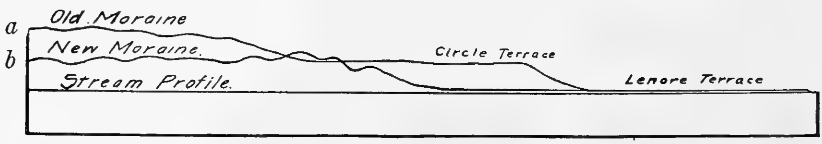

FIG. 38.—Diagram showing the relation of Circle and Lenore terraces to moraines of the (*a*) Bull Lake, and (*b*) Pinedale glacial stages. Wind River basin.

On the whole, the Lenore cycle did not go far, and resulted only in the excavation of relatively narrow valleys with slender and even interrupted floodplains, except in weak strata and along large streams. The moraines of the latest glacial stage were deposited in these valleys, showing definitely that the inner trenches were made just before the last ice advance.

It will some day be possible, through detailed local studies, to subdivide the Lenore cycle on the basis of terraces, shallow inner trenches, etc. For example, the Snake River in Jackson Hole is intrenched distinctly below the general level of the flat; at the mouth of the Grand Canyon of the Snake, the river has intrenched its gravel-strewn floodplain to a depth of 50–60 feet; and near the mouth of the Hoback River that stream has cut a trench of about the same depth through deposits of gravel into solid rock below. These features are so much more prominent in the Snake River drainage than in the Green River or Wind River basins that I am

led to think that the three systems have had somewhat different histories. Along Wind River and Owl Creek the streams are now generally intrenched 10–30 feet below the gravel-strewn terrace of the Lenore cycle, and widened according to circumstances. This has been accomplished in post-glacial time.

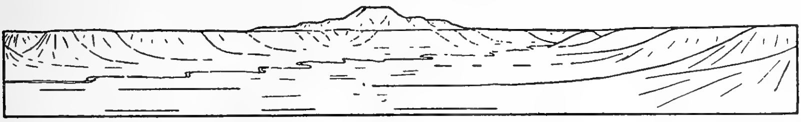

FIG. 39.—Crow Creek valley and Crowheart Butte, showing the broad Circle plain, beneath which the modern valley has been cut, surmounted by a monadnock carrying a remnant of an older plain. (Drawn from a photograph.)

FIG. 40.—Dry Creek valley in the Wind River basin. It illustrates the dissection, in the Lenore cycle, of the plain of Circle. The beds dip 10–15° away from the camera. The valley in the center of the view is about 150 feet deep. (Photographed by D. D. Condit, U.S. Geological Survey.)

Quaternary glaciation.—All of the mountain ranges of the district that exceed 10,000 feet in elevation bear marks left by valley glaciers of an earlier time. Indeed, a few small bodies of ice still exist in the deep cirques of the Teton, Shoshoni, and Wind River ranges. By glacial action the mountain crests have been excavated in the characteristic way, and the canyons have been modified

in varying degree. In some cases the ice descended as far as the foothills of the range, and a few of the largest ice-tongues expanded out upon the adjacent plains. It is not my purpose to map or describe in detail the glacial phenomena of each range, but rather to discuss the general nature and course of events in the district as a whole, using individual cases for illustrations and as evidence.

It is clear that the time which has elapsed since glaciers first appeared in this region has been very long, and the time since they vanished very short. On the one hand, there are deposits of till that have been so deeply eroded that only remnants of them now

FIG. 41.—The floodplain of Fish Creek in the Mount Leidy highlands. Benches on the left are remnants of the older (Circle) floodplains, cut across inclined beds.

remain, and the intervening canyons are hundreds of feet deep in solid rock. On the other hand, some of the glacial features are as fresh as if they had been made within historic times—the moraines almost untouched by erosion, the bowlders essentially unweathered, the lakes not yet filled, and the striated rock surfaces still conspicuous. In the older drift little if any trace of original glacial topography remains, and most of the surface bowlders have disappeared either through weathering or through burial by eolian dust. An examination of only one or two valleys in the district might prompt the conclusion that there were two and only two stages of

glaciation; but, after having studied the deposits of more than forty-five Pleistocene glaciers, distributed over a large part of western Wyoming, I find it no easy task to reduce the facts to such comfortable simplicity. Indeed, if disposed to support arbitrarily the hypothesis that there have been many successive advances of glaciers, I could, by selecting cases, arrange a graded series leading by almost imperceptible steps from the least modified glacial deposits which have been practically unaffected by post-glacial changes, to those which have lost most of their glacial characteristics and have been eroded to isolated remnants. Before discussion of the question of successive glacial stages some general considerations need attention.

The glacial features in the several valleys differ from each other in many respects, among which the following are important: (*a*) the depth to which moraines have been intrenched by streams, and the extent to which they have been reduced in area; (*b*) the degree of preservation of distinctive glacial topography; (*c*) the closeness of relation of the moraines to modern valleys and their terraces; (*d*) the progress of weathering of glaciated rock surfaces and of the bowlders upon the moraines; (*e*) the depth of soil, particularly of loess, on the moraines.

These attributes are influenced by more or less variable conditions and one should be prepared to find drift deposits of essentially the same age, changed much in some cases and little in others. Lateral moraines resting upon soft clays may be rapidly destroyed through the growth of gullies, while moraines planted firmly on the hard Paleozoic or pre-Cambrian rocks suffer only trivial erosion in the same length of time. Drift plastered along steep slopes may be largely denuded while contemporaneous moraines lying upon interstream plateaus retain their original topography. Ponds without inlets apparently are filled much more slowly than those entered by streams, especially if those streams are eroding soft materials. Where the climate is dry, as in the Wind River basin, it is probable that the bowlders on the moraines would show fewer signs of decay than those in the moist forests on the west slope of the Teton Mountains, and also that soil and loess would accumulate more rapidly in the latter region,

because the vegetation serves to entrap the dust from the western basins.

This prelude to the general discussion of Quaternary glaciation is given in order to show that the satisfactory correlation of the different drift deposits and other glacial phenomena of the district is a difficult matter, calling for critical interpretation. In order to eliminate the effect of variable conditions, it is necessary so far as possible to compare only cases where the environment has been generally similar; and I may add that a large amount of additional information from the field is needed.

In many canyons glacial drift of two distinct ages can be recognized easily, and it seems to me there is sufficient evidence that the later stage of glaciation is divisible into two quite distinct episodes. There are a few facts which hint at a still older or fourth stage, but they may be otherwise interpreted. Without attempting an actual correlation, the three sets of deposits may be compared in general state of preservation to the late and early Wisconsin, and the Illinoian or the Kansas drift sheets in the Mississippi Valley region. In explanation of this view I shall describe the characteristics of the drift and other glacial features of the various stages here suggested, beginning with the most recent.

Pinedale stage: The youngest moraines, which for convenience may be called the Pinedale drift, are conspicuous in the vicinity of Torrey Lake and Bull Lake in the Wind River valley, Osborn ranch on Green River, Jenny Lake and Taggart Lake in Jackson Hole, around each of the large lakes near Pinedale on the southwest side of the Wind River Range, and also in many other valleys of the district. They are very rough and fresh in appearance, being covered with bowlders which themselves show but little effect of weathering (Fig. 42). Many of the bowlders of hard, fine-grained rocks show polished or scratched surfaces, even though fully exposed to the weather; and in the cirques above the moraines such surfaces, bare or now covered only with moss, are abundant. The terminal moraines are generally intact and the lateral moraines continuous, save where unusual conditions have subjected them to abnormally rapid erosion. Lakes and ponds which contain but little post-glacial filling are characteristic features of these newest

moraines, and even those lakes which lie in the courses of the main creeks have been only slightly reduced by the building of deltas or the cutting of outlets.

Bull Lake stage: The next older moraines, which I have compared with the early Wisconsin drift of Illinois, will be called the Bull Lake drift, from the locality of that name on the north slope of the Wind River Range. The deposits of this stage are definitely related to the terraces of the Circle erosion cycle in that they blend

FIG. 42.—Hills in the latest terminal moraine at the lower end of Bull Lake. The prominence of bowlders is characteristic.

into the gravel-strewn benches as modern moraines blend into their outwash deposits. The moraines are still largely intact, and by using a little imagination to supply gaps, they can be mapped in their original extent. In protected places, as on flat uplands, the topography is still visibly morainic, consisting of orderless humps and hollows. That the Bull Lake drift is nevertheless distinctly older than the Pinedale drift seems to me to be indicated by the following facts: (*a*) bowlders are by no means as abundant upon the

surface as they are on the Pinedale moraines, and those which remain show plainly the effects of weathering;[1] (*b*) lakes are rare or absent, but hollows filled with peat mark the sites of many that have

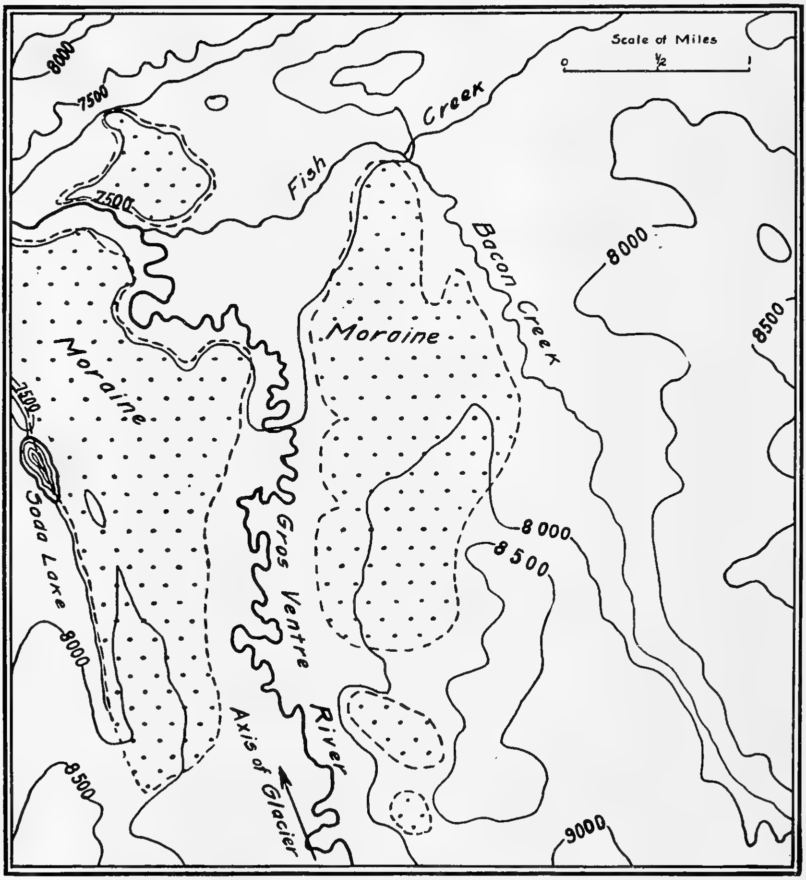

FIG. 43.—Sketch map showing the dissection of the old moraine (Bull Lake [?] stage) at the junction of Fish Creek and the Gros Ventre River.

disappeared; (*c*) the principal creeks have cut not merely sharp ravines but rather wide flat-bottomed valleys through the terminal

[1] A fact possibly related to the weathered condition of the older drift is the observed scarcity of cedar trees on the older moraines at Bull Lake, while such trees are so abundant on the adjacent new moraines that the latter appear much darker.

moraines, and the tributaries have largely dissected the lateral moraines into short pieces.

The frayed character of the terminal moraines is indicated in Figs. 43 and 44. At the mouth of the DuNoir River near the west

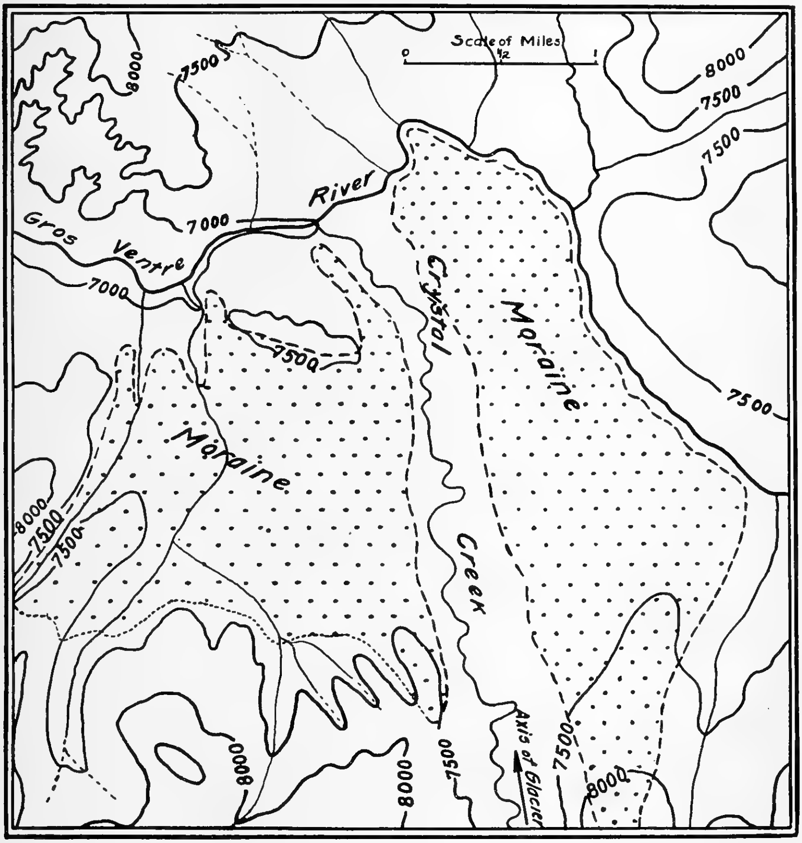

FIG. 44.—Sketch map showing the dissection of the old moraine (Bull Lake [?] stage) at the mouth of Crystal Creek, Gros Ventre Range.

end of the Wind River basin, the position of the old terminal moraine is occupied by a spacious river floodplain; nevertheless, the old lateral moraine, now 200–300 feet above the river, is still largely intact and can be readily identified by its hummocky topography. On the north side of Dell Creek, on the south side

of the Gros Ventre Range, two well-defined lateral moraines, without any terminal junction, have been beheaded upstream by the excavation of the inner valley of the modern Shoal Creek. On account of the much more rapid destruction of lateral moraines where they rest on steep slopes in soft rocks, the old lateral moraine at Sunday Peak on the Gros Ventre River has been almost completely destroyed by the development of numerous short ravines in the Cretaceous shale.

Some of the best examples of the Bull Lake moraines are to be found near the mouth of Crystal Creek (Fig. 43), and near the mouth of Fish Creek (Fig. 44) in the Mount Leidy quadrangle, along Green River south of Kendall in the Gros Ventre quadrangle, and the older moraines of Bull Lake, DuNoir River, and Dinwoody Creek in the Wind River basin. At the mouth of Teton canyon, on the west slope of the Teton Range, distinct lateral moraines have been cut into short lengths and yet they are still connected downstream by the eroded remains of a bulbous terminal moraine. On the basis of similarity of conditions these deposits are correlated with the moraines of the Bull Lake glacial stage.

The glaciers of the Bull Lake epoch were in most instances longer than those of the Pinedale epoch. Thus, on Crystal Creek the difference in length was about four miles, on the Gros Ventre River about nine miles, and on Green River apparently twenty-five miles. On the north slope of the Wind River Range, however, there seems to have been but little difference in length, and the Pinedale moraines in the valleys of Bull Lake and Dinwoody Creeks were even advanced a fraction of a mile farther than the older terminal moraines.

Buffalo stage: Certain other glacial deposits, which are evidently very much older, are readily differentiated. These may be called the Buffalo drift, from their typical occurrence along Buffalo Fork of Snake River. The deposits which appear to represent this stage now exist only in the form of remnants on flat-topped divides or isolated hills, or on spurs along valley slopes. Such surfaces have already (p. 312) been correlated as belonging to the Black Rock cycle of stream erosion. The grouping of these remnants suggests that they are parts of originally more extensive bodies

which have since been cut to pieces by the growth of canyons. The canyons have been excavated, not only through the drift sheet itself, but 200–1,000 feet in the underlying rock. Thus the deep valley of the north fork of Teton River has been cut almost entirely since the deposition of the Buffalo drift, for the river now has a V-shaped gorge devoid of glacial features in the horizontal flows of rhyolite, and the old drift caps a flat-topped divide on either side (Fig. 45). The canyon of Buffalo Fork, about fifteen miles above its junction with Snake River, is a similar trench sunk about 1,000 feet into the Cretaceous (?) shales which underlie the extensive deposits of old drift.

As should be expected, the ancient moraines have been much more extensively destroyed where they rest upon weak strata in proximity to an actively degrading stream. For example, in the

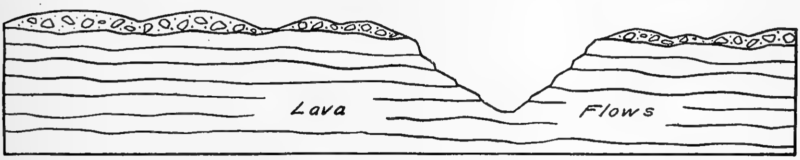

Fig. 45.—Diagram of the canyon of Teton River southeast of Ashton, Idaho, showing the present relations of the oldest (Buffalo) drift.

valley of the Gros Ventre River about seventeen miles above its mouth, there is a remnant of till identified by striated subangular bowlders exposed in a new roadway. The granitic and other distinctive rocks entering into the composition of this material indicate that the ice-tongue came originally from the Wind River Range, and probably from the canyon of Green River. Yet no trace has thus far been found of the old lateral moraines.

Uncorrelated drift deposits greatly reduced by erosion were observed on the high divide between Leigh and Teton creeks, on the west slope of the Teton Range. There the glacial origin of two or three little bodies of bowlder-clay is proven by the presence of striated stones, and other characteristics of till, but their age is not clear. One of the remnants occupies a saddle in an undulating ridge which separates two deep canyons. Another remains on the shoulder of a spur dividing the forks of a canyon 800 feet deep.

It is evident from their situations that they were deposited before the adjacent canyons were cut. It is also patent that they are but the last relics of the original moraines, of which the outlines and catchment areas can no longer be seen.

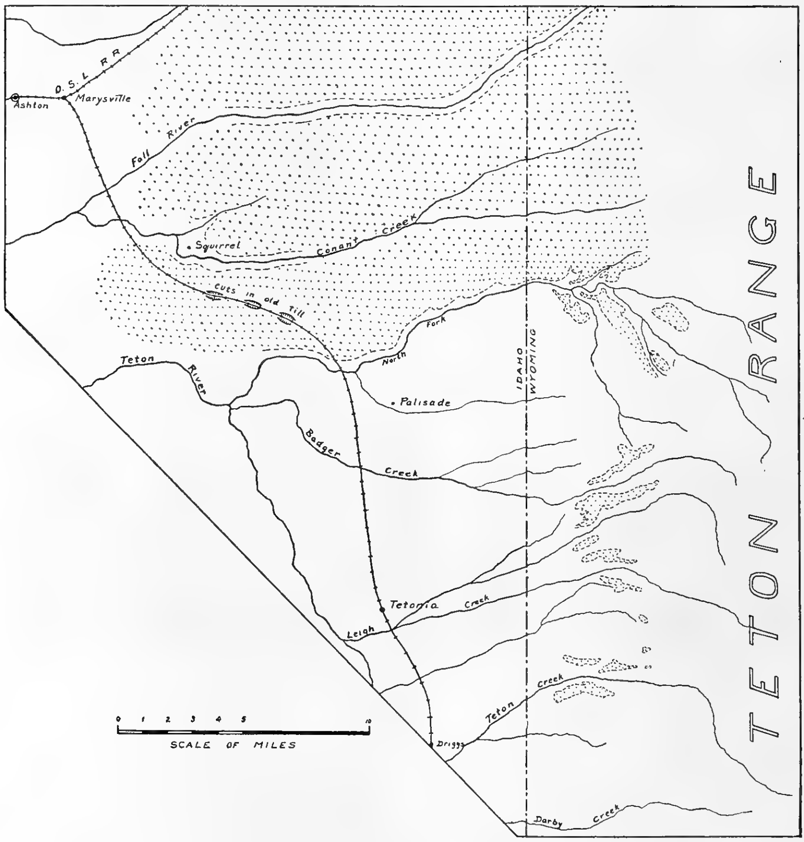

FIG. 46.—Sketch map showing distribution of remnants of old (Buffalo) drift on the northwest side of the Teton Range.

They may be regarded as drift deposited by a small glacier of the Buffalo stage from the slope between Leigh and Teton creeks, and subsequently almost destroyed by the spreading tributaries of the larger canyons. It is also possible that they may represent an earlier glaciation.

Cirques of the Buffalo stage are no longer to be identified with confidence, and it is now uncertain whether the glaciers were of the valley type or parts of an ice-cap. In most cases even the general outline of the ice-mass is uncertain. But few traces of morainic topography now remain in the drift of this age, because most parts of the region have been dissected to maturity by the growth of valley-systems. This condition is clearly shown on the broad sheet of till along the new branch of the Oregon Short Line between Marysville and the north fork of Teton River in eastern Idaho (Fig. 47). Nevertheless, on some of the broadest plateaus, where

FIG. 47.—The present topography of the ancient till sheet (Buffalo) south of Conant Creek on the northwest slope of the Teton Range.

relief is still slight, there are shallow depressions filled with peat, or still containing marshy lakes. Such features are to be found twenty miles east of Ashton, Idaho, and near the branching of Buffalo Fork of Snake River.

The surface of the old drift in the western half of the region is generally smooth and almost devoid of bowlders. This probably is due not so much to the decay of bowlders as to their burial by the eolian deposition of loess in that district to a depth of from two to five feet. Bowlders of granite on the isolated hills of sedimentary rock south and east of Kendall in the Green River valley are outside

the zone of loess deposition, and are now so much weathered that the large feldspar crystals protrude and some of them can be picked out with the fingers. The coarse, unassorted, and unstratified character of these deposits suggests that they may be glacial; and their isolated situation indicates that they have suffered greatly from erosion. They are tentatively referred to the Buffalo drift.

Glacial scratches are of course found only on bowlders of resistant rock and even then only where they have been effectually insulated from the action of the weather. Striated bowlders of Buffalo age were found in considerable numbers in the new railroad cuts, southeast of Ashton, Idaho. In that locality the old drift is more or less cemented by lime carbonate.

From these facts it seems certain that the changes which have taken place since the deposition of these bodies of till are very much greater than those which occurred between the making of the Bull Lake and Pinedale series of moraines. The conditions imply that the period between the Buffalo and the Bull Lake glacial stages was several times as long as the next succeeding interglacial interval, and many times longer than the post-glacial epoch.

Glacial drift comparable to the Buffalo moraine in its relations seems to cover most of the plateau of Yellowstone Park. Even in the seventies, as a member of the Hayden Survey, Holmes[1] recognized the fact that the deep gorge of the Yellowstone must have been cut after the deposition of the glacial drift which mantles the surrounding plateau. For some unaccountable reason, this early recognition of drift much older than the fresh moraines in the Rocky Mountain region seems to have remained unnoticed and is not mentioned even in more recent comprehensive papers on the geology of Yellowstone Park itself.

Although its full extent is no longer traceable, it is evident that the Buffalo ice covered a much larger area than the glaciers of the Bull Lake and Pinedale epoch. The latter were valley glaciers, which in but few instances pushed beyond the mouths of the canyons in which they developed. The Buffalo ice of the Teton back-slope, and probably also of Yellowstone Park, spread far

[1] W. H. Holmes, "Glacial Phenomena in the Yellowstone Park," *Am. Naturalist*, XV (1881), 203–8.

out upon plateau surfaces at least twenty-five miles from the mountains.

Summary of glaciation: From this discussion a few generalizations arise. All of the principal ranges of the district, except perhaps the Owl Creek Mountains, were glaciated during the Quaternary period. There were certainly two stages of glaciation separated by a long period of time, and the later of these can again be resolved with some confidence into two. A few facts slightly suggestive of a fourth epoch, still older than the others, are quite inconclusive. The glaciers of the Pinedale and Bull Lake stages were ordinary valley glaciers, which ran but little beyond the mountain fronts and rarely deployed outside their individual canyons. In most valleys, the younger of these glaciers were shorter than the older, although in a few instances they were not. In the Buffalo stage, however, a large piedmont glacier spread westward from the north end of the Teton Range, and Yellowstone Park appears to have been almost entirely buried under a mass of ice.

There is much resemblance between the sequence of glacial events in this district and that described by Atwood in the San Juan Mountains of Colorado. Although full confidence should not be placed in a long-distance correlation made without indulging in a more detailed and comprehensive study of the Rocky Mountains, it may be suggested that the deposits characteristic of Atwood's "Uinta," "Bighorn," and "San Juan" stages correspond in many respects to those of the Pinedale, Bull Lake, and Buffalo stages in western Wyoming. The Pinedale and "Uinta" moraines are alike in being very fresh and almost uneroded. The Bull Lake moraines resemble somewhat those of Atwood's "Bighorn" stage in being notably eroded although still traceable, but the Bull Lake moraines are rather better preserved. Like the till of Atwood's "San Juan" stage, the Buffalo moraines have been intrenched 500–1,000 feet and so reduced in area that at best their original forms can be but rudely traced, and in some places they constitute mere remnants which suggest nothing of former distribution or derivation.

Work of gravity with the aid of weathering.—Here, as in all regions of sharp relief, the steeper slopes are wasting away under the attacks of the weather; and the resulting débris has been making

its way down the slopes by rolling, slumping, and creeping. No doubt these processes have all been in operation, at varying rates, throughout the erosion history of the district. There is but little recognizable evidence of them, however, relating to topographic stages farther back than the Black Rock cycle of erosion.

FIG. 48.—Talus glacier in a cirque on the north side of Sheep Mountain, near the head of Green River.

The accumulations of talus are greatest above the timber-line, and beneath the most resistant cliff-forming strata. In the lower parts of glaciated valleys, the waste slopes have been largely swept away, except in so far as they have begun to redevelop since the ice disappeared. In a few cirques, talus glaciers, such as those described by Cross and Howe[1] in the San Juan Mountains of Colorado, have

[1] *U.S. Geol. Survey, Prof. Paper 67*, 1909.

been formed. On the northeast side of Sheep Mountain at the northwest end of the Wind River Range there is a good example showing the tongue-like form and surface corrugations (Fig. 48). Rock-falls have been observed where outcrops of the massive Bighorn dolomite have given way suddenly, perhaps because of the removal of the support previously given by the mass of glacial ice in the valley. Figs. 49 and 50 show two rock-falls of this kind in the Teton and Gros Ventre ranges. More numerous landslides of the commoner type, developing in soft or unconsolidated rocks, are observable here and there along the slopes of the canyons in the

FIG. 49.—A rock fall in Paleozoic dolomite near the head of the Gros Ventre River

district. There is a particularly large and fresh slide near the head of Dell Creek on the south side of the Gros Ventre Range. Perhaps the most novel of the many landslides in this region are the earth-flows which are especially characteristic of the Gros Ventre valley (Fig. 51). They depend upon steep slopes in soft argillaceous material such as constitutes the Jurassic, Cretaceous, and Tertiary systems. The most recent of these flows I have described in another paper,[1] and for this reason further mention of them may be omitted here.

The landslides and talus deposits are clearly of various ages. A few have been made within recent years. Many others are

[1] *Bull. Geol. Soc. Am.*, XXIII (1912), 487–92.

covered with forest and must be some generations or centuries old, although nearly all of their topographic features and forms are still preserved. Still others have been scarred by gullies to such a degree that they are now recognizable only upon somewhat careful examination. There are many places in which landslides of two or three distinct ages may be seen side by side on the same valley slope.

Wind work in the Quaternary period.—The climate of the entire region is moderately dry, and in the centers of the large basins it is decidedly arid. Under such circumstances the wind is admittedly

Fig. 50.—A rock-fall in the canyon of Darby Creek, Teton Range

an important agent in shaping topographic forms. The characteristics of its work are, however, greatly influenced by the varying conditions in different parts of the district. Thus, in the hot, dry central part of the Wind River and Green River basins, the erosive action of the winds can be observed daily. Bunches of sagebrush and greasewood are left perched on hillocks, where the silt has been blown out from around them by the wind. Protruding ledges of rock have been carved by the sand-blast into the usual fantastic pillars, mushroom forms, and honeycomb surfaces. Many of the smooth slopes along the sides of the valleys are swept so clean that every lamina of rock may be measured or traced for yards and even miles along the outcrop. The smooth tops of many mesas and

terraces out in the bad lands coincide with the more massive and hence resistant layers of sandstone from which the wind has completely swept off the soft beds of sandy clay.

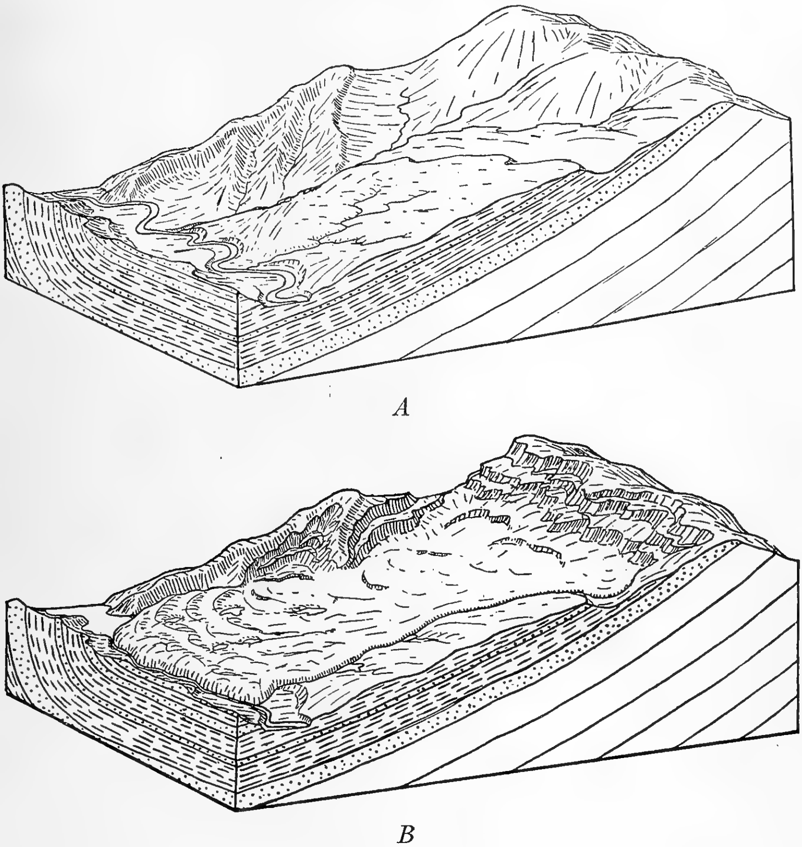

FIG. 51.—Diagrams to illustrate the formation of an earth-flow such as the "Gros Ventre Slide" of 1909–11: (*A*) before the action began; (*B*) after the slide took place. (From *Bull. Geol. Soc. Am.*)

On the other hand, wind erosion must necessarily have wrought but slight effects upon those parts of the mountain sides that are protected by forests, and but little more upon their rocky summits. The timber belt, constituting Merriam's[1] Hudsonian and Canadian

[1] C. H. Merriam, "Life Zones and Crop Zones of the United States," *Bull. 10, U.S. Geol. Survey*, 1894.

life zones, and the sagebrush and aspen slopes of his Transition zone, are apparently, however, the sites of considerable eolian deposition. For example, along the grassy and forested western slope of the Teton Range, a thick structureless soil mantles the surface. Occasional exposures show that it is from two to twenty feet deep, entirely free from sand, gravel, or bowlders, and of a yellowish or buff color. The chemical analysis given in Table III shows that it is similar in composition to the loess of the Mississippi and Missouri valleys.

TABLE III

TABLE OF ANALYSES OF LOESS*

	1 Teton Mountains, Wyoming	2 Iowa	3 Missouri
SiO_2	72.04	72.68	74.46
Al_2O_3	12.37†	12.03	12.26
Fe_2O_3	3.38	3.53	3.25
FeO	0.37	0.96	0.12
TiO_2	....	0.72	0.14
P_2O_5	....	0.23	0.09
MnO	....	0.06	0.02
CaO	1.21	1.59	1.69
MgO	1.22	1.11	1.12
Na_2O	1.83	1.68	1.43
K_2O	2.58	2.13	1.83
H_2O	3.15	2.50‡	2.70
CO_2	trace	0.39	0.49
SO_3	....	0.51	0.06
C, organic	....	0.09	0.12
Undetermined	1.85	....	....
	100.00	100.21	99.78

* Analyses 2 and 3 are taken from a paper on "The Driftless Area" by T. C. Chamberlin and R. D. Salisbury, *6th Ann. Rept., U.S. Geol. Survey*, 1885, p. 282.

† Including TiO_2, P_2O_5, MnO, etc.

‡ Contains H of organic matter. Dried at 100° C.

1. Loess overlying old drift, one mile east of Alta, Teton Creek, Teton Mountains, Wyoming. Sample taken two feet below the surface. Analyst, W. C. Wheeler, U.S. Geol. Survey, 1912.

2. Loess near Dubuque, Iowa.

3. Loess from river bluff at Kansas City, Missouri.

It is suggested that this widespread veneer of loess consists largely of the dust blown by the westerly winds from the dry regions of Oregon and Idaho and strained out by the vegetation on the mountain slopes. It is significant that such deposits are not characteristic of the Wind River and Green River basins, which

are apparently subject to the removal of dust rather than to its accumulation.

Sand dunes, which might be expected in the driest parts of the district, appear to be rare, although a few were observed along the middle course of Green River. This tends to confirm the view that in this region the dominant phase of eolian activity is erosive.

SUMMARY OF CONCLUSIONS

Divested of details and qualifications, the history of the district may now be reviewed in general terms. About the close of the Cretaceous period, the entire region was probably a monotonous plain underlain by unconsolidated sediments. Soon thereafter, tangential forces within the earth's crust produced a series of moderate folds, trending northwest and southeast. The process of folding tended to produce a series of mountain ranges and basins corresponding to the anticlines and synclines; but the uplifts were certainly being eroded even while they were growing, and after the disturbance ceased they continued to be worn down until the surface became a well-dissected mountainous or hilly region interspersed with wide, relatively flat basins.

In the Wasatch or Lower Eocene epoch, terrestrial sediments began to accumulate in the lowlands, probably in response to mild warping movements and perhaps climatic changes, which disturbed the activities of the previously graded streams. These sediments were distributed chiefly by rivers, but doubtless in part by the wind, and some seem to have lodged in lakes and playas. Sedimentation thus continued with but little interruption under a rather warm, subarid continental climate until some time after the early Oligocene. Early in this sedimentary cycle there were volcanic eruptions in the western part of the district, during which acidic lava flows and occasional beds of ash were deposited; but the much more extensive eruptions of the Absaroka Range took place later. The grassy plains of the time were occupied by ancestral horses, cloven-hoofed ungulates, rodents, gigantic titanotheres, and their appropriate carnivorous enemies.

Somewhat later, probably near the middle of the Miocene epoch, the district was notably warped and locally faulted, in part

along the structural lines established at the close of the Cretaceous, but in part at variance with them. It is probable that this mid-Tertiary disturbance involved also a general elevation which left the region above base-level, and therefore subject chiefly to erosion rather than to sedimentation.

From the mid-Tertiary revolution down to the present time, the history of western Wyoming is a chronicle of denudation. It is believed that at all times the master streams of the region controlled the general shaping of topography, although the effects of stream erosion were modified in important details by mountain glaciers, by the wind, and by slumping. The progress of general stream erosion was evidently not steady, but was interrupted by successive elevatory movements or other changes which compelled the readjustment of stream activities. In the earliest of the erosion cycles of which good evidence still remains, the region seems to have been generally reduced to a peneplain even where the rocks were tolerably resistant. This plain is believed to have been established during the Pliocene epoch.

Later cycles followed each other with more rapidity—or more probably we read the later chapters in more minute detail. They were apparently caused in large measure by general uplifts of the Rocky Mountain region, although they may well have been influenced in noteworthy degree by climatic changes, river piracy, and other modifying causes. While these erosion cycles and interruptions were in progress, there occurred certainly two, and almost surely three, distinct advances of glacial ice in the mountains, followed by as many retreats and perhaps complete disappearances. Between the glacial stages the streams continued their work and sank their valleys through the moraines and into the underlying rocks. Since the latest glacial advance only slight topographic changes have been made.

THE GEOLOGY OF CENTRAL MINAS GERAES, BRAZIL

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PART I

INTRODUCTION

The South American continent bears resemblance to the North American continent in several respects. Like the northern continent it may be divided according to its geologic history into an eastern half and a western half. Like North America its Atlantic

side was deformed and thrown into mountains before the close of the Paleozoic. As in our own Appalachians this mountain-building period of eastern South America was followed by long-continued erosion which base-leveled the mountains. Later successive widespread uplifts produced low plateaus, out of which rejuvenated erosion has carved the present ranges. Aside from these gentle uplifts, which have occurred at several periods during later geologic history, the eastern part of the continent has exhibited comparative stability for a long time. Since the Paleozoic the deformation of South America has largely been on the Pacific side. During the Mesozoic and Tertiary remarkable stressed conditions on the Pacific border developed the great world-ridge of Cordilleras which runs from Cape Horn to Bering Strait and which makes the western portions of the two continents also geologically related.

The area considered in the present paper lies in the eastern half of the South American continent. It is that section of the state of Minas Geraes, Brazil, which lies between Diamantina (latitude 18°17′ S.) and Ouro Preto, the former capital of the state (latitude 20°23′ S.), which is situated east of the 45th meridian. In many respects this area is geologically the most typical and complex part of eastern Brazil. Running north and south across it is the Serra do Espinhaço, or Backbone Ridge of Brazil, which forms the eastern rim of the São Francisco basin and divides the waters flowing into the Rio São Francisco from those which flow directly into the Atlantic Ocean. Structurally and historically this mountain range gives the key to the geology of most of eastern Brazil, for either within the range, or closely associated with it, are all the important sedimentary formations of Minas Geraes. These comprise a great series of quartzites, schists, and iron formations with some associated limestones. East of the Serra do Espinhaço is a great area of dominantly igneous rocks, presumably Archean, which extends, with the exception of small isolated patches of overlying sedimentary rock, continuously to the sea. West of the Serra do Espinhaço is a plateau country underlain by rocks of both the sedimentary series and the basement complex.

Within this region structurally so significant, occur also the most important mineral deposits in Brazil (Fig. 1). Here is

found the gold which caused this portion of the country to be settled at an early date. Here also are found the diamonds and other precious stones for which Brazil has long been famous. And it is along the flanks of this strip of the Serra do Espinhaço that there

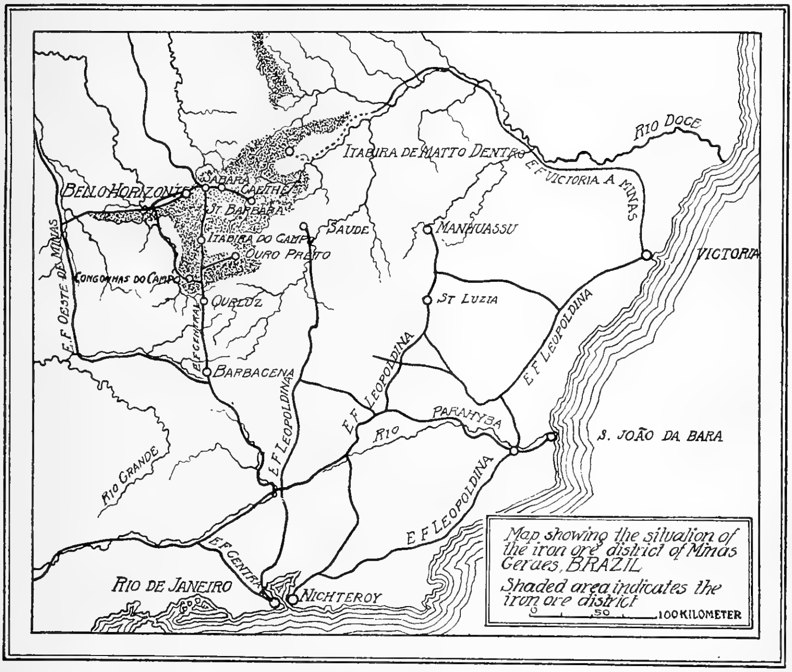

FIG. 1.—Map showing the location of the iron-ore district of Minas Geraes, Brazil

are being discovered today the still more remarkable deposits of high-grade iron ore which seem destined to develop a great industry in the future.

GENERAL GEOLOGY

GENERAL DISTRIBUTION, SUCCESSION, AND STRUCTURE OF FORMATIONS

The coastal region of Brazil throughout the states of Rio de Janeiro and Espirito Santo is composed, except for small areas of Mesozoic or Tertiary sediments occurring locally along the ocean border, of a complex of ancient crystalline rocks of which granite, gneiss, and crystalline schist are the prevailing types, and in which diorite, gabbro, and other more basic rocks occur as

intrusives. These rocks form the basement complex of eastern Brazil and are probably to be referred to the Archean. The crystalline belt extends far into the interior, at first continuously, but farther away from the coast, as patches and arms amid areas of overlying post-Archean sedimentary formations.

Eastern Brazil has been, throughout much of its geologic history, a region more subject to denudation than to sedimentation. In Minas Geraes the uncovered basement complex is now areally the most extensive surface formation. The basal granite, gneiss, and crystalline schist are nearly continuous throughout the eastern and central portions of the state. Where the younger sedimentary beds do occur, they persist in favored localities as local remnants of former widespread formations. The post-Archean sedimentary series is best represented within the limits of the Serra do Espinhaço which owes its existence and present elevation to the superior resistance of some of its quartzite beds.

The area of post-Archean sedimentary rocks of central Minas Geraes extends in a continuous belt of varying width from about latitude 20°40′ S., in south-central Minas Geraes, northwestward across the state and into the state of Bahia. Southward from latitude 20°40′ S. isolated areas of sediments continue into southern Minas. As a result of complex earth movements, the southern end of the principal sedimentary belt is wide and irregular, occupying nearly all of the area from a point about 15 kilometers southeast of Ouro Preto, northwestward to Bello Horizonte, a width of about 80 kilometers. Followed northeastward the belt narrows but continues as a strip from 5 to 20 kilometers in width almost to Diamantina, where it branches, one limb running northeastward into northeastern Minas Geraes and the other continuing northward into Bahia. Both on the east and west, this strip of sediments is bounded by extensive areas of older rocks, probably Archean, consisting of gneiss, granite, and crystalline schists. Areas of sediments, of unknown age, however, occur at various places west of the sedimentary belt, becoming more extensive toward the north.

The metamorphosed sediments, which Derby has called the Minas series,[1] are probably of Algonkian age, judging from their

[1] O. A. Derby, "The Serra do Espinhaço, Brazil," *Jour Geol.*, XIV (1906), 396.

stratigraphic position and lithology and their similarity to Algonkian sediments in other places. They lie upon the eroded surface of the basement complex, and have as their basal member a great quartzite formation. Above this quartzite, often separated from it by a thin argillaceous schist bed, is the peculiar iron-bearing quartzite known as itabirite, accompanying which are extensive iron-ore lenses. This formation is, in turn, overlain by a schist and quartzite formation of great thickness. The uppermost known member of the great series is a quartzite with associated schists. Occurring in proximity to these metamorphosed sediments, and derived from them by disintegration and decomposition, are subaerial deposits of more recent age.

Appropriate geological names taken from localities where characteristic exposures of the various formations occur have been adopted. The known succession is as follows:

Tertiary and Quaternary
- River gravels
- Tertiary clay and lignite
- Canga deposits

Mesozoic or Early Tertiary
- Diamantina conglomerate

Probable Algonkian
- Itacolumi quartzite
- Piracicaba schist and quartzite
- Itabira iron formation
- Batatal schist
- Caraça quartzite

Probable Archean
- Gneiss, granite, and schist

The various sedimentary formations are complexly distributed, occurring in some places and being absent elsewhere. Their irregularity of occurrence is due, in part, to irregularity in original deposition, but more especially to later faulting and folding followed by very extensive erosion. They are separated from the basement complex by a profound unconformity.

The most characteristic formation of the series is the Caraça quartzite which extends throughout the length and width of the sedimentary area and locally develops an extraordinary thickness.

Somewhat more limited in occurrence is the iron formation which is well developed in the broad portion of the area between Ouro Preto and Bello Horizonte, but which to the northeast and southwest becomes inconspicuous or is entirely absent. The upper formations are also local in their occurrence (Fig. 2).

The Caraça quartzite forms the most prominent mountain system of this portion of Brazil which has taken the name Serra do Espinhaço, or Backbone Range, and which divides the waters flowing westward into Rio São Francisco from those flowing eastward directly into the Atlantic Ocean. The iron formation generally forms well-marked foothills along the quartzite ridges, though locally, where well developed, and specially hard, it itself forms the main ridges while the quartzite is on the slopes. The Piracicaba schist, being soft, is inconspicuous topographically, but the Itacolumi quartzite forms a number of prominent peaks and ridges.

The distribution of formations is well shown by the topography. The mountain peaks and long continuous ridges common to this region are made up of quartzite and iron formation. They are bounded by lower areas of schist, or by an irregular undulating region of hills and valleys underlain by gneiss and granite. These sedimentary rock mountains of central Minas Geraes are, with the exception of the granite masses forming the Itatiaya Range on the southern border of the state and Caparaó Mountain on the boundary between Minas Geraes and Espirito Santo, the most conspicuous mountains in Brazil.

The characteristic deformation of the rocks of central Minas Geraes has been thrust faulting and accompanying folding. The forces in general came from the east and south, ranging between northeast, southeast, and south, forces from different directions being applied at different times, resulting in a sort of superimposition of structure. This complex structure is especially characteristic of the wide area between Ouro Preto and Bello Horizonte, where the combination of forces has resulted in the formation of a number of parallel and intersecting belts of sedimentary rock. In the southern part of this area is an east-west syncline resulting in two parallel belts of sediments; in the western part is a north-south overturned syncline causing a similar distribution, and in the

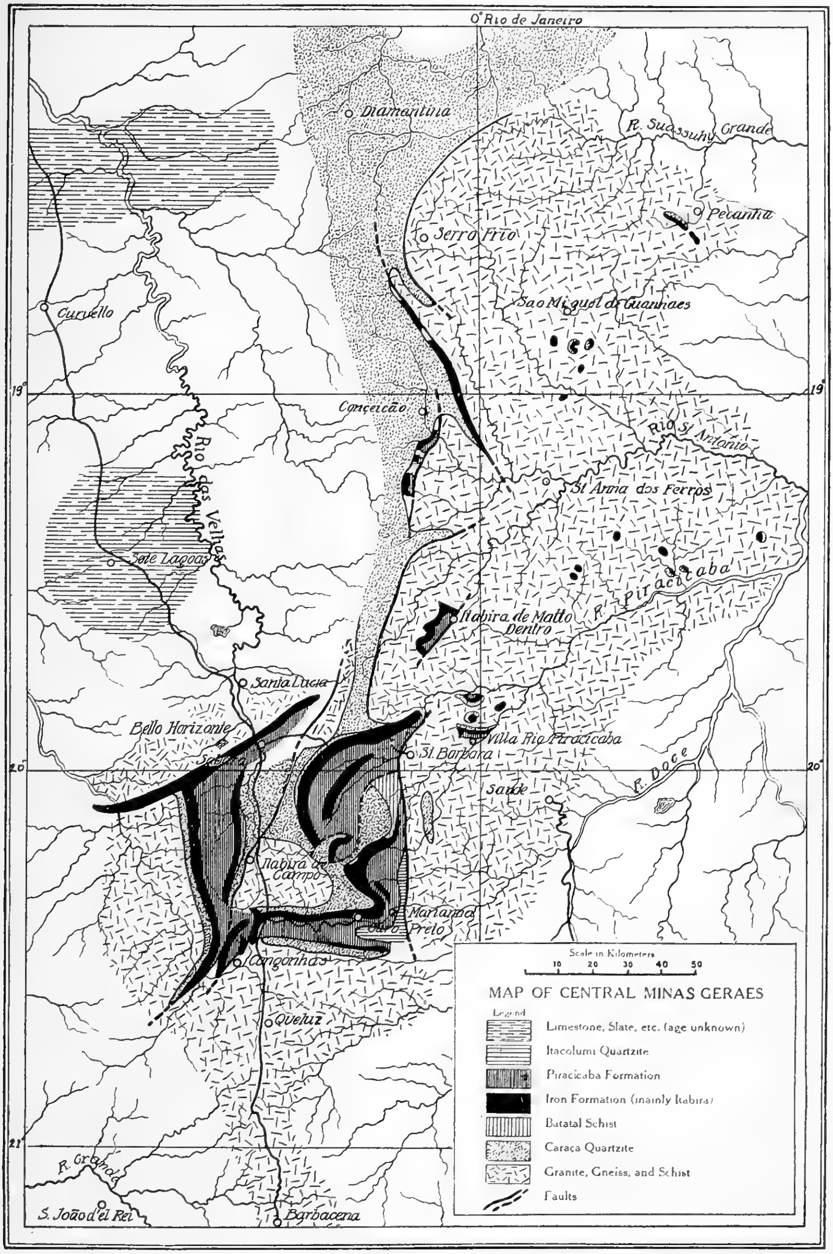

FIG. 2.—Map of central Minas Geraes

northern and eastern parts is a series of more or less parallel thrust faults, resulting in several parallel and branching belts of sediments. In the center, occupying the angle between the two synclines, is a large circular area of granite, gneiss, and schist of the basement complex raised by this combination of forces into a dome structure.

A very conspicuous result of the deformation has been the development of V-shaped branchings of the formations in various parts of the area, such as those south of Capanema, west of Cocaes, southwest of Bello Horizonte, and northwest of Serro. These V-shaped junctions can be explained by forces working in different directions at the same or at different times, in such manner that, when breaks occur in the formation, one portion of it is forced against another portion at an angle. Frequently at such junctions there has been much crushing and dislocation of the formations.

In general the irregularity of deposition, resulting in a great variation of thickness as well as of materials, combined with the complex structural relations above mentioned, has resulted in a condition of complexity of structure and areal distribution of formations not easy to solve.

PROBABLE ARCHEAN

Basement complex.—The rocks of the basement complex consist chiefly of granite of varying composition and texture, gneiss with interlayered amphibolite, and micaceous and quartzose schists. Granite and gneiss are probably the most abundant of the rocks in the Archean areas, and of these two the granite is the more commonly observed because of its greater hardness. It forms prominent ridges in many parts of central Minas, and is commonly exposed in rounded bluffs along streams where these flow through granite areas. Crystalline schists are also abundant but are less conspicuous because, on account of the softness of these rocks, exposures are not so common. These schists resemble closely the schist beds in the Piracicaba formation and sometimes can be distinguished from them only by their general distribution. Elsewhere they may be distinguished by their more prominent recrystallization, or by the presence of pegmatite veins. Such

pegmatite veins are also abundant in the gneiss but are rarely seen in the granite. They usually consist of muscovite, orthoclase, and quartz.

The amphibolite is a banded rock occurring interlayered in the gneiss and corresponding in layering to the gneiss itself. It is a dark-green rock consisting either entirely of hornblende, or of hornblende with a small amount of feldspar. It is widely distributed throughout the region of the basement complex, and in many localities masses of considerable thickness occupy large areas. On weathering the amphibolite gives rise to a yellow ocherous soil due to the formation of abundant iron oxide. Near the surface this changes in color to a deep red which is very characteristic of the soils in many parts of the district.

Diorite and gabbro occur sparingly as later intrusions in the gneiss and schist in different parts of the district, diorite being more common than gabbro. Most of the intrusions are in the form of dikes, some of which may be traced for long distances. Some of the gabbro intrusions, however, are very limited in extent.

The relation of the three principal varieties of rock in the basement complex to each other is determined with difficulty on account of the complexity of their structure and distribution, and because of the thick covering of mantle rock which effectively conceals the greater portion of these rocks. The granite is apparently intrusive into the gneiss and schist and occurs in large irregular areas. Single masses 60 or 70 kilometers in extent are known. The granite intrusions are probably older than the gabbro and diorite. The relation between the schist and gneiss is unknown. In places the schist has been seen interlayered with the gneiss and apparently bears a relation to it similar to that which the amphibolite bears to the gneiss. If this is true the schist, gneiss, and amphibolite would seem to be the older rocks into which granite, diorite, and gabbro were intruded. Probably the gneiss and amphibolite represent respectively ancient acid and basic volcanic flows with which were interbedded quartzose and argillaceous sediments now represented by the schists. By metamorphism and deformation these were transformed into their present equivalents—

the igneous rocks to gneiss and amphibolites, the sediments to crystalline schists. The later intrusions may have taken place during, or after, the deformation.

Many areas of schist, the definite structural relations of which it has not been possible to determine, have been presumed to be part of the basement complex. During further investigations certain of these schists may be found to belong to one, or other, of the overlying sedimentary formations.

Lithologically the schists of the basement complex are extremely various. There are quartz-muscovite schists, sericite schists, talc schists, chlorite schists, amphibole schists, and argillaceous mica schists of varying texture, many of them ferruginous, others siliceous. The gneiss, on the other hand, is fairly regular in texture and composition throughout most of the district. In most localities it is medium grained with distinct banding, though elsewhere the banding is so faintly developed that it is difficult, by lithology alone, to distinguish the gneiss from later granitic intrusives. Its predominant minerals are feldspar, quartz, biotite, and hornblende.

From the foregoing descriptions it is clear that to distinguish between the various rocks of the basement complex, to separate them areally and determine their structure and relations to each other, will require much petrographic study and detailed mapping.

POST-ARCHEAN EROSION PERIOD

During the deformation and metamorphism of the gneiss, schist, and amphibolite, and their intrusion by the granite and other igneous rocks, the surface of the central Minas Geraes region was probably of a very rugged and mountainous character, and was probably elevated far above the surface of the sea. Then a period of erosion and disintegration began, for a long time the former being predominant, but later as the mountains were being worn down and acquired gentler slopes disintegration became more and more important. For a long time the products of disintegration and decomposition were largely carried away to the sea, but gradually, as the slopes became more gentle, material began to accumulate and a layer or blanket of soil formed over the solid rock. This

thickened as erosion and transportation decreased and as it thickened it became more and more thoroughly decomposed.

While there were still considerable differences of elevation the sea began to encroach upon the land, gradually occupying what is now eastern, southern, and central Minas Geraes. Apparently an arm of the sea extended westward covering the southern end of Goyaz, the southern and western part of Matto Grosso, and probably part of Bolivia. Whether the sea extended far southward into São Paulo is not known, but it is doubtful, since sediments of this age are unknown in central and southern São Paulo. To the north the sea covered northeastern Minas Geraes and a large part of Bahia, but whether it extended still farther north is not known.

It was in this sea that the probable Algonkian sediments of central Brazil were deposited. Owing to the great amount of decomposition which the rocks of the basement complex had undergone previous to the encroachment of the sea, little remained in the soil except the end products, especially kaolin, iron oxide, and silica. As the end products were derived largely from acidic granites and gneisses, silica dominated. In the shallow advancing sea the waves may be supposed to have been capable of sifting these residual materials, holding in suspension and carrying away the fine mud and iron oxide while of necessity soon depositing the coarser grains of quartz. The first material laid down upon the eroded surface of the Archean complex in the area under discussion was therefore a great mass of quartz sand. By cementation, induration, and metamorphism this has become the Caraça quartzite formation.

PROBABLE ALGONKIAN

Caraça quartzite.—The Caraça quartzite is the basal formation of the thick sedimentary series of supposed Algonkian age. Over most of the region the formation consists of more or less schistose quartzite, though it embraces a complete range from pure vitreous quartzite to quartzite schist made up almost entirely of white mica. The variations in composition and texture occur along the beds, as well as across them, so that at different places along the quartzite belts the major part of the formation may be either predominantly schistose or predominantly quartzitic. Locally the

formation may be thin-bedded, while elsewhere masses many meters in thickness and of great uniformity in texture may occur without a trace of bedding.

The distribution and structure of the quartzite are determined by the general structure of the district (Fig. 3). The dips are in general to the east or southeast at varying angles, except for minor irregularities such as occur in the region between Ouro Preto and Congonhas do Campo. Even where three or four parallel or slightly intersecting belts of quartzite occur, this regularity of

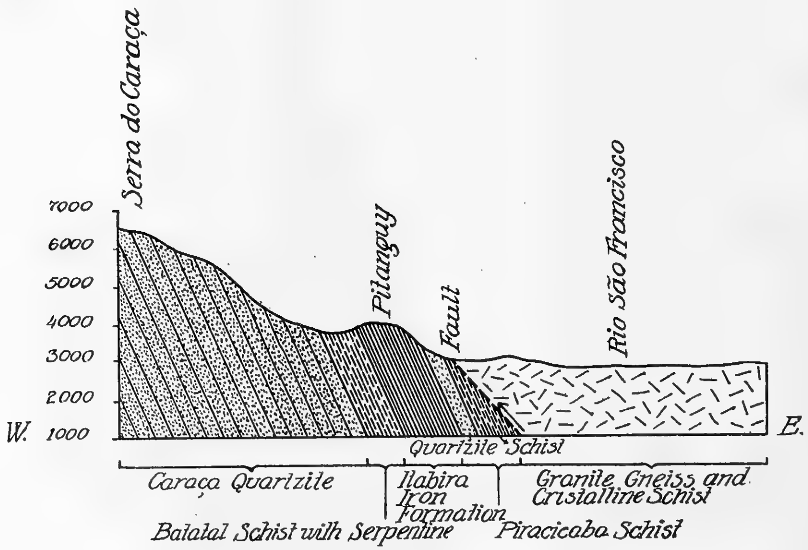

FIG. 3.—East front of the Serra do Caraça. Section through Pitanguy Hill. Elevations given in feet.

dip prevails. Such parallel belts of southeasterly dipping strata are the results of thrust faulting and overturn folding, and their regularity indicates the uniform nature of the deformation to which the region has been subjected.

The Caraça quartzite though varying in thickness extends throughout the length of the sedimentary belt, being everywhere well developed except in the region east and south of Bello Horizonte where it appears to be absent, for the iron formation lies directly on the basement complex. Along considerable portions of the

belt, especially in the northern part, it is the only one of the Algonkian formations present, being bounded on both the east and the west by rocks of the basement complex. In detail the distribution of the Caraça quartzite is as follows: Beginning at the southwest, a belt of quartzite dipping to the southeast runs northeastward to a point about northwest of Congonhas do Campo where it turns northward and continues with great uniformity to about the latitude of Bello Horizonte where it is cut off sharply by a northeast-southwest belt of sediments. To the east of it, and approximately parallel with it, is a shorter quartzite belt also dipping to the east. Between the two belts are areas of iron formation and Piracicaba schist, while to the east and west of them are granite, gneiss, and schist of the basement complex. From the structure it is presumed that these two north- and south-striking quartzite belts are on opposite limbs of an overturned syncline dipping to the east. Both belts are well shown in the topography, the western one appearing as the Serra da Boa Morte and the eastern one forming the high hills west of Itabira do Campo.

Beginning with Serra do Ouro Branco near Miguel Burnier is an east-west belt of quartzite somewhat more irregular than are the north-south belts, being broken at several places, and varying in strike in different parts, from northwest-southeast to northeast-southwest. It continues eastward to a point southeast of Ouro Preto where it is cut off by rocks of the basement complex. South of it, unconformably underlying it, are schist, gneiss, and granite of the basement complex. Some distance to the north of it and parallel to it is another quartzite belt dipping southward, being underlain and bounded on the north by granite, gneiss, and schist. Between these two quartzite bands are areas of iron formation and Piracicaba schist, so that here occurs an east-west syncline differing from the north-south syncline before mentioned in being broad and open. The southern limb forms the prominent Serra do Ouro Branco while the northern limb is not very marked topographically, being mainly composed of quartzite schist not very resistant to erosion.

The northern of these quartzite belts continues westward beyond Miguel Burnier and eastward to Ouro Preto where it

makes a sharp turn northward and then continues northwestward, thus forming a narrow eastward pitching anticline. The northern limb of the anticline dips to the northeast and rocks of the basement complex are exposed southwest of it along the axis of the anticline. At Capanema the quartzite band branches, one arm going northeastward and forming the conspicuous mountain mass of the Serra do Caraça (Fig. 4) while the other continues northwestward, forming

Fig. 4.—The Serra do Caraça at Alegria. A portion of the Backbone Range. The high Serra consists of resistant Caraça quartzite. In the middle distance are hills of the Itabira iron formation whose strata dipping toward the right reach a maximum thickness of 1,200 meters in this section. In the far distance to the right are lowlands of the granite complex.

the Serra da Capanema, then, becoming less prominent topographically, it continues northward around the headwaters of the Rio Santa Barbara.

At a point west of Cocaes the latter belt divides again, an eastern arm forming the prominent ridge above Cocaes and then disappearing, while the main belt here forms the central axis of the Serra do Espinhaço, which continues northward for many miles—a great quartzite range.

Several minor offshoots to the east occur from the northward extension of the range, and at a point west of Serro the range divides, one branch running northeastward, forming the Serra do Chifre, on the divide between Rio Doce and Rio Jequitinhonha, and the other continuing northward past Diamantina toward the state of Bahia.

The northeast-southwest belt of sediments previously mentioned as cutting off the north-south overturned syncline near Bello Horizonte consists mainly of Itabira iron formation and Piracicaba schist dipping southeast. Locally a thin layer of argillaceous and quartzitic schist, representing the Caraça quartzite, occurs at the base. The belt is of considerable extent, beginning at a point northeast of Caethé, east of the main Serra do Espinhaço, and running first southwestward past Bello Horizonte and the end of the north-south overturned syncline already mentioned, and then westward and northwestward toward Pitanguy. It forms a prominent range known as the Serra da Piedade which consists for the most part of iron formation. Toward the west, however, the quartzite becomes more prominent, forming some conspicuous mountains.

The Caraça quartzite, while present almost throughout the district, varies greatly in thickness. In the isolated sedimentary areas in the eastern part of the district, as near Villa Rio Piracicaba, there are places where it is less than 30 meters thick; in the region south and east of Bello Horizonte it appears to be in places entirely absent, while along the main Serra do Espinhaço it reaches thicknesses greater than 1,500 or 1,800 meters. The latter is the more typical development of this great formation.

The ordinary phase of the Caraça quartzite consists of quartz grains intermixed with more or less white mica. Where highly metamorphosed, and this is especially true where the formation is thin, the mica is sometimes coarsely crystalline so that individual flakes frequently reach half an inch in diameter. When quartz occurs in these phases it is also coarsely crystalline. Such coarsely crystalline phases, however, are rare. Where the formation consists largely of quartzite, it is prominent topographically, but where the schist layers are abundant and well developed, it gives way rapidly to erosion.

Both the quartzite and the schist phases are light colored, the quartzite being white, light green, or light brown according to the stage of oxidation or decomposition of its constituents. The light-brown color is due to the oxidation of ferrous iron present and is prevalent near the surface. The schists are usually white, light green, or light gray in color owing to the principal constituent, white mica.

Batatal schist.—After the deposition of a great thickness of the coarse clastic sediment which makes up the Caraça quartzite, there came a change in the character of the material being washed into the sea. The sediments became finer, whether owing to the lowering of the land surface from which the material was derived, or to a further advance of the sea upon the land. Whatever the cause, there occurred a slackening of sedimentation as well as progressive fineness. The result was the laying down of a series of varicolored clays which through metamorphism have resulted in the Batatal formation, a light-gray to dark-red, fine-grained, argillaceous schist, which overlies the Caraça quartzite conformably. This schist is perhaps best represented along the base of the Serra do Caraça, though, on account of its softness, it is seldom exposed in outcrop. Not only inconspicuous in outcrop, it is also a comparatively thin formation, seldom exceeding 30 meters in thickness. At one point, however, along the base of the Serra do Caraça northwest of the village of Cattas Altas, it suddenly thickens to more than 300 meters of schist, while at the same time the overlying iron formation gradually pinches out. The situation is somewhat peculiar and it is probable that the upper portion of the Caraça quartzite at that point is schistose, and thus is indistinguishable from the Batatal schist.

On the eastern and southeastern flank of the Serra do Caraça a sheet of serpentinized eruptive rock is found to rest upon the Batatal schist at a number of points. It is seen at Morro da Mina, and is also conspicuous in Boa Vista, near Cattas Altas, extending several kilometers northward from there till cut off by a fault. This serpentine layer always occurs at the same horizon at the top of the Batatal schist and at the base of the iron formation. It is not an intrusive, but was apparently a sheet of basic lava

spread out upon the then level surface of the Batatal sediments. The iron oxide and silica deposits which now constitute the Itabira iron formation were then laid in regular succession upon it. If this basic flow were more widely extended throughout the iron district and were of greater mass, one might be inclined to look to it for a source of the iron now locked up in the great iron formation which followed so closely upon it. Such an explanation has been advanced to explain the origin of much of the iron formation in the Lake Superior region. In their great monograph on the geology of the Lake Superior region, Van Hise and Leith have come to the conclusion that the iron formations there are so intimately asso-

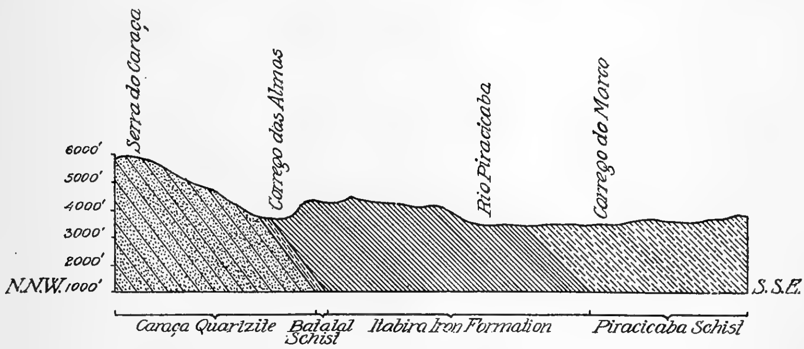

Fig. 5.—Section through Fazenda da Alegria

ciated with basic lavas that the source of the iron in the great ore beds is to be sought in the igneous activity.[1]

In Minas Geraes, however, remnants of a basic lava flow near the iron formation in the stratigraphic column have been discovered up to the present time only at the two adjacent localities just mentioned. Though these probably belong to a single flow which extended continuously from Morro da Mina nearly to Santa Barbara, this flow after all covered but a comparatively limited area which, taken together with its thinness, makes it quite insignificant in size and extent in comparison with the tremendous iron formation above it (Fig. 5). Thus it seems more reasonable to account for the deposition of the iron formation through some other agency.

[1] C. R. Van Hise and C. K. Leith, "The Geology of the Lake Superior Region," *Mon. 52, U.S. Geol. Surv.* (1911), pp. 409–570.

Itabira iron formation.—The Batatal schist represents a slackening of sedimentation from the rapid deposition which characterized the laying down of the sands composing the Caraça quartzite. This slackening of clastic sedimentation continued until the close of the Batatal epoch, when very little clastic material was being washed into the sea in the region considered. The land presumably had become so low as to yield very little mechanical sediment, and with this lowering of the land surface there was probably combined a gradual retreat of the shoreline. Simultaneously with the great diminution in mechanical sediment deposited in the area under consideration, there commenced a precipitation of ferric hydroxide from solution, materials in solution being probably carried beyond the borders of the region of clastic sedimentation. This precipitation may have been due, either to purely chemical reactions taking place in the sea, or perhaps to the operation of the well-known iron bacteria which cause the deposition of ferric hydroxide from waters containing ferrous carbonate in solution. These iron bacteria are said to possess the peculiar property of utilizing as food the carbon dioxide locked up in very dilute solutions of ferrous carbonate.[1] Ferric hydroxide is left behind and is deposited as a sediment. The characteristic reaction may be written: $2FeCO_3+3H_2O+O=2Fe(OH)_3+2CO_2$. This process is operative in very dilute solutions. Apparently only two or three parts of iron per million are needed to make certain types of the iron bacteria active.

Not having much confidence in the hypothesis that the iron oxide was precipitated directly from sea-water by ordinary chemical means, we prefer to turn to the iron bacteria as perhaps forming a better working hypothesis. If such a process as this be supposed to have been steadily in operation at this time, when very little clastic sediment was being deposited, it might have resulted in the production of the extensive iron formation. For reasons which will be discussed later, this hypothesis seems to explain the

[1] S. Winogradski, "Über Eisenbakterien," *Botan. Zeitung*, Bd. 46 (1888), p. 261; Rudolf Lieske, "Beiträge zur Kenntnis der Physiologie von Spirophyllum ferrugineum Ellis, einem typischen Eisenbakterium," *Jahrb. für wissenschaftlicher Botanik*, Bd. 79 (1911), pp. 91–127.

observed facts connected with the formation of the Itabira iron beds better than any other yet suggested, and to encounter fewer serious difficulties.

Like the schist and quartzite, the Itabira iron formation is a true primary sediment.[1] It is a mixture of iron oxide and quartz sand laid down essentially as it occurs today. The principal change which it has undergone since its deposition is the dehydration which has converted the ferric hydroxide originally laid down into ferric oxide, or hematite. This dehydration, like the metamorphism of the shale into Batatal schist and the sandstone into Caraça quartzite, probably is to be associated with the intense deformation which these strata suffered during the mountain-forming movements which antedate the Devonian.

The Itabira iron formation takes its name from Itabira Peak, near the town of Itabira do Campo, a rather striking mountain of splendid specular hematite which forms a conspicuous landmark visible for many miles around. Though varying greatly in character, the Itabira formation is in general hard and resistant and much more durable than the softer schists which lie immediately above and below it. The result is that wherever the iron formation now appears at the surface in inclined beds, it stands up in prominent ridges or as a chain of "iron hills." In some places where both the iron formation and the underlying Batatal schist are very thin, the former may occur simply as a capping, or cover sheet, following the contour of ridges of Caraça quartzite. Elsewhere the iron formation constitutes a separate range of hills in front of, and parallel to, the quartzite ridge. The hills of iron formation are as a rule lower and less massive than those of quartzite, though locally, as along the western rim of the upper Santa Barbara valley, where the quartzite is relatively soft and contains a good proportion of schistose layers, the iron hills overtop the quartzites.

The iron formation varies greatly in thickness. In the ridge running east from Morro Agudo near Villa Rio Piracicaba, the whole

[1] C. K. Leith and E. C. Harder, "The Hematite Ores of Brazil and a Comparison with the Hematite Ores of Lake Superior," *Econ. Geol.*, VI (1911), 670–86; E. C. Harder, "The 'Itabirite' Iron Ores of Brazil," *ibid.*, IX (1914), 101–11.

formation ranges between 5 and 20 meters in thickness. In Morro Agudo Peak it suddenly thickens to 100 meters (Fig. 6). Thirty-odd kilometers away to the southwest, in the locality known as Alegria, on the east slope of the Serra do Caraça, it has a thickness of over 1,200 meters, which is probably its maximum. It is not a case of steady increase, or decrease, in thickness between areas some distance apart, for at the very eastern end of Alegria the iron formation is comparatively thin, and the swelling from a few

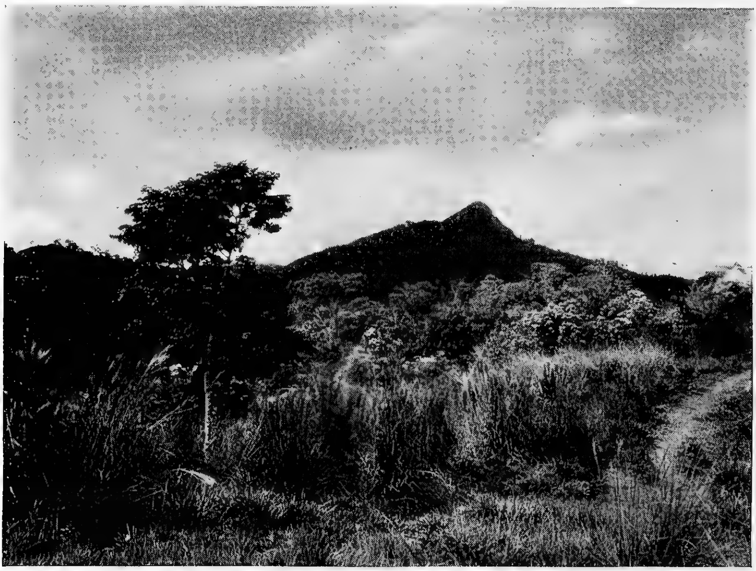

FIG. 6.—Morro Agudo, near Villa Rio Piracicaba. A peak of high-grade iron ore which owes its form to the superior resistance of the hard, specular hematite of which it is composed.

score up to 1,200 meters is accomplished in the amazingly short distance of about 3,700 meters along the strike (see Fig. 4). This very remarkable thickening of the deposit in such a short distance strongly suggests the conditions of delta deposition. If the layers of iron formation represent the foreset beds laid down, shingle fashion, on a delta front, the great total thickness of the successive layers may be readily understood, and in this way some of the grave difficulties which would otherwise be encountered in the development

of such a thick deposit are avoided. In the eastern part of the district, as along the Serra da Boa Morte, there seems to be much more uniformity in the thickness of the formation.

In chemical composition the iron formation also varies greatly, and the changes are as sudden as the variations in thickness. In some places, the entire thickness of the formation may be almost pure iron oxide containing less than 1 per cent of impurity. At other points there may have been so much arenaceous material incorporated within the iron sediments that the proportion of metallic iron is reduced to 40 per cent, or even less. But whether the percentage of iron be high or low, the impurities consist largely of rounded grains of quartz sand together with a little phosphorus, the mineralogical relations of which are not known. In places, however, the iron oxide is mixed with clay rather than sand, and beds of ferruginous schist are interlayered with sandy iron formation. Thus, while ferric hydroxide was being precipitated, a certain amount of sandy sediment, and locally, also, clay were washed in to mingle with the precipitate. The presence of impure limestone beds in the lower part of the iron formation, as southeast of Bello Horizonte, shows that carbonate precipitation locally accompanied the iron-oxide precipitation. In some portions of the formation, the sand is more or less evenly disseminated, while in others it occurs more often as distinct sandy partings, or laminae, which separate thin beds of relatively pure iron oxide. The temporary inwash of clastic sediment in the form of sandy partings probably marks unusual storms, or stormy periods, during which more of the terrigenous material found its way beyond the borders of general clastic sedimentation. The rock breaks more readily along the planes of these sandy partings than elsewhere, and hence possesses a somewhat sandy fracture face, so that it appears to contain more sand than is actually the case.

The purer portions of the iron formation, where the proportion of metallic iron exceeds 50 per cent, are regarded as iron ore, while the more sandy portions of the formation are designated itabirite. Formerly the term itabirite was applied as a formation name to the entire iron formation member of the sedimentary series, and included the iron ore as well as the sandy portions and schistose

layers. But now its use is being restricted, and the name is applied only in a petrographic sense to the peculiar iron-oxide bearing quartzite which constitutes the sandy portion of the iron formation.[1] There is, however, a complete gradation between the itabirite and the ore.

Piracicaba formation.—After a period of very little wash from the land, during which the iron formation was laid down through rapid precipitation chiefly, an epoch of more abundant clastic sedimentation again set in. This change was probably brought about by a gentle uplifting of the land relative to the sea-level. At first only the finer muds were laid down in the region under consideration; later, with changing conditions, there were oscillations both in the direction of diminishing clastic sedimentation and a return to the iron-oxide and calcium-carbonate forming conditions, and also in the direction of increasingly coarse clastic sedimentation with the deposition of what has since become quartzite. Under these fluctuating conditions which followed the deposition of the Itabira iron formation there was laid down a great mass of sediment, in which clays and muds predominated, but within which there were included quite extensive lenses and more or less irregular beds of quartz sand, iron oxide, and also calcium-carbonate deposits. Through subsequent metamorphism these sediments were altered to argillaceous and quartzite schist, quartzite, iron formation, and limestone. The name Piracicaba formation has been given to this series from the Piracicaba River, whose upper course near Santa Rita Durão and in Alegria follows the strike of these beds.

The lenses and beds of limestone and iron formation occur principally in the lower part of the Piracicaba formation, being interlayered with schists, while the quartzite beds predominate in the upper part. Limestone lenses, though usually small, are numerous in many parts of the district, especially in the central and southwestern parts. The limestone is generally impure, containing, besides calcium carbonate, carbonates of magnesium, iron,

[1] The term itabirite is coming into more general usage as is shown by the fact that it has been applied to iron-oxide bearing quartzites occurring in certain parts of Europe. See F. Beyschlag, J. H. L. Vogt, P. Krusch, *Ore Deposits*, I, 113, 194 (translation by S. J. Truscott).

and manganese. Different lenses differ in composition, some being reddish or purplish in color and containing manganese carbonate as the principal impurity, while others are grayish or brownish and contain iron carbonate as impurity. In places thin layers of itabirite occur in the limestone lenses, or considerable thicknesses of interlayered limestone and itabirite may occur.

The iron formation lenses in the Piracicaba formation are generally more extensive than the limestone lenses but are more local in their distribution. In lithology they are similar to the Itabira iron formation, consisting mainly of itabirite but containing also lenses of pure hematite and beds of ferruginous schist. In most cases the iron formation lenses stand up as ridges above the inclosing schist.

The schists making up the lower part of the Piracicaba formation are mainly argillaceous, gray, red, or purple in color. In the upper part sericitic, talcose, and quartzose schists are more common.

Next to the Caraça quartzite the Piracicaba formation is probably the most widespread of the metamorphosed sedimentary rocks. It is almost coextensive with the former, except in the northern part of the district where it is absent in many places on account of faulting, but on the other hand it is present in places where the Caraça quartzite is absent, as in the region southeast of Bello Horizonte.

Like the other sedimentary rocks, the Piracicaba formation is of variable thickness. Furthermore it is not always easy to draw the division line between this formation and the overlying Itacolumi quartzite, since beds of quartzite become increasingly prominent, in the upper portion of the Piracicaba formation. And as the Itacolumi quartzite is represented only at a comparatively few points in the district, the full thickness of the Piracicaba formation is to be observed only in a limited number of places. But judging from the exposures where the entire formation is present, it would seem to be seldom less than 300 meters in thickness and generally many times this. It is a great mass of sediment, of sufficient bulk to lay a heavy tax upon nearly any system of dynamics and method of sedimentation which may be supposed to have resulted in this thick pile of beds.

Itacolumi quartzite.—The series of indurated rock formations is capped, so far as the iron-ore district is concerned, by a thick quartzite which is not very different lithologically from the Caraça quartzite. As this upper formation is perhaps displayed to best advantage in the prominent peak of Itacolumi, the popular lookout point near Ouro Preto, it may well take its name from that well-known mountain. It is the youngest member of the regular sedimentary series yet discovered in central Minas.

While this formation where best exposed is predominantly quartzite, it contains also a large proportion of schistose rocks. The quartzite grades laterally into sericite, quartz, and talc schists, so that, in places, the formation is largely composed of schist with a few prominent and persistent layers of quartzite. This variation in lithology is so characteristic both of the Piracicaba formation and of the Itacolumi quartzite that the two may perhaps well be regarded as one formation, being formed during a continuous period of deposition, schists predominating in the lower part and quartzite in the upper part. They strongly suggest delta deposition, individual layers being seen to grade from coarse quartzite into argillaceous schist within a distance of 4 or 5 kilometers.

In fact, there is much to indicate delta deposition in most of the members of the Minas series. The Caraça quartzite, Itabira iron formation, Piracicaba formation, and Itacolumi quartzite all exhibit rapid lithological variations as their layers are traced along the strike as well as across it. At the same time, they show also in some places great thicknesses and extremely rapid variations in thickness. These are not the results to expect if the sediments were laid down in regular succession upon the relatively flat and uniform ocean floor; but they are just what might be anticipated if they represent the topset, foreset, and bottom-set beds of delta formation. Under the former hypothesis it is not easy to see how such great thicknesses of shallow water deposits could be formed under any rational system of crustal dynamics; under the hypothesis of inclined sedimentation the total thickness of the series of shingling layers laid down in shallow water may very easily become as great as the strata of the Minas series require.

The name Itacolumi is an old one in geological literature and has been variously applied by different writers. As far back as 1822 Eschwege proposed the name itacolumite for a quartzose talco-micaceous rock which he found abundant in, and characteristic of, the gold and diamond fields of the Serra do Espinhaço.[1] Eschwege himself recognized that the rocks to which he gave the name itacolumite were not a unit, and in consequence noted two groups, one schistose and the other massive, but he lumped together under the general name of itacolumite all the quartzose rocks of the region, which has been an incubus on Brazilian geological studies.[2]

For a long time these rocks were generally regarded as belonging to the primitive igneous crust of the earth,[3] but later geologists have shown them to be clearly metamorphosed sediments. By the more recent writers (Derby, Gorceix) the two divisions of the original itacolumite have been considered to be independent formations.[4]

While the term itacolumite has been variously applied to the quartzites and quartz schists of the Serra do Espinhaço, it also has been used as a petrographical term to designate the peculiar flexible sandstone, or quartzite, which is an oddity found at several points in the district.[5] But this property of flexibility is characteristic of only an insignificant portion of the whole formation.[6] If the term be retained for the future, its only proper place would seem to be as the petrographic name for this flexible phase of the quartzite, for there seems to be little reason for applying a single mineralogical or petrological term to such a great and varied succession of metamorphosed sedimentary formations as are now found to make up the Backbone Range.

[1] Wilhelm von Eschwege, cited by O. A. Derby, "On the Accessory Elements of Itacolumite, and the Secondary Enlargement of Tourmaline," *Am. Jour. Sci.*, V (1898), pp. 187–92.

[2] O. A. Derby, "The Serra do Espinhaço, Brazil," *Jour. Geol.*, XIV (1906), footnote, pp. 374–75.

[3] A. de Lapparent, *Traite de Geologie*, 2d ed., 1885, p. 654.

[4] O. A. Derby, *Am. Jour. Sci.*, V (1898), 187.

[5] E. S. Dana, *Textbook of Mineralogy*, p. 190.

[6] O. A. Derby, "On the Flexibility of Itacolumite," *Am. Jour. Sci.*, 3d ser., XXVIII (1884), 203–5.

Itacolumite as a formation name has been used, not only for the quartzite above the iron formation (the quartzite out of which the peak of Itacolumi has been carved), but also for the great quartzite series (Caraça quartzite) which, throughout the district, underlies the iron formation. That the quartzite which makes up the peak of Itacolumi is distinct from the tremendous quartzite formation of the Serra do Caraça is a fact which does not seem to have been brought out clearly by those who have studied this region; nor has it been clearly recognized that the quartzite which forms such a large part of the Serra do Espinhaço from Ouro Preto northward to Diamantina is the lower. But these two quartzite formations are quite distinct and are separated from one another by the Itabira iron formation and two schist series. If the name of Itacolumi be applied to one of the quartzite formations, it would seem fitting that it should be applied to that one which occurs in the peak of Itacolumi. This is the upper quartzite.

In areal extent the Itacolumi quartzite is much more limited than the Caraça quartzite. This is chiefly the result of erosion which has swept away the upper members of the Minas series from many localities where the lower members still remain. The Itacolumi formation is nowhere better displayed than just south of the city of Ouro Preto in the mountain ridge which culminates in the peak of Itacolumi, where there are perhaps 1,200–1,500 meters of quartzite exposed. The upper quartzite appears again on the north limb of the Ouro Preto anticline, as a conspicuous ridge just north of the village of Bento Rodrigues.

Whether the Itacolumi quartzite was once overlain by younger sediments of the same general series, or not, is uncertain. As yet, no evidence of a younger member of the series has been recognized. But as the Itacolumi quartzite, though a resistant formation, remains only at a few points in the whole district, it is not impossible that younger, and less resistant, beds may once have capped it and since been removed by erosion.

Summary.—This whole great series of sedimentary formations is apparently conformable throughout. Above the base of the Caraça quartzite no unconformity of any moment has been noted. But from top to bottom, not a single fossil has been found within

the series which might aid in determining its geologic horizon. The strata have been metamorphosed to such an extent that fossil forms might originally have been present in these beds and have since been obliterated in the process of metamorphism. But to what extent the absence of fossils is due to an original paucity of impressions in the original sediments, and to what extent to obliteration of existing forms by subsequent metamorphism we cannot say. The age of the series therefore remains problematic. It is simply known to rest unconformably upon the probable Archean complex. The general nature of the strata harmonizes well with the hypothesis that these beds are of pre-Cambrian age. This is the age generally assigned to the series by the geologists of the Brazilian Survey. At the present time we can hardly go farther than to note the general parallelism between this Brazilian metamorphic series and the Proterozoic sediments of other parts of the world, especially the Lake Superior Algonkian series. The heavy quartzites, schists, and iron formations indicate not very widely different conditions in these two regions. But while the sedimentary series in Minas presents a strange parallelism to the Algonkian series of the Lake Superior region in North America, and to various other pre-Cambrian terranes, no definite evidence has yet been found to preclude its belonging in whole, or in part, to the early Paleozoic.[1]

PRE-DEVONIAN DEFORMATION

Just when this long period of sedimentation in central Minas was interrupted is not yet apparent. What is known is that it was followed, whether shortly, or after a considerable interval, by a period of great deformation. Mountain-making movements profoundly affected a belt running across central Minas.[2] The strata were both folded and thrust-faulted on an extensive scale. At the same time the rocks suffered metamorphic changes; the sandstones became quartzites, the shales were altered to the present schists, the carbonate rocks were recrystallized, and the iron formation dehydrated to the hematite and itabirite which we see today.

[1] See J. C. Branner, *Geologia Elementar*, Rio de Janeiro, 1906, p. 217.

[2] O. A. Derby, "The Serro do Espinhaço, Brazil," *Jour. Geol.*, XIV (1906), 347–401.

The rocks did not everywhere yield to the stresses in the same way. In the southeastern part of the district the strata were thrown into open folds, though these folds show less symmetry than those of the Jura or the Appalachian system. With the folding there is also some faulting associated. But in the northeastern portion of the district very little folding has occurred. That region has suffered, instead, very extensive thrust faulting which repeats the sedimentary series, and by thus repeating the outcrops of iron formation becomes a factor of much economic importance. The fractures have occurred along lines trending for the most part northeast and southwest, or parallel to the Atlantic coast. The faults are overthrusts from the southeast and east.

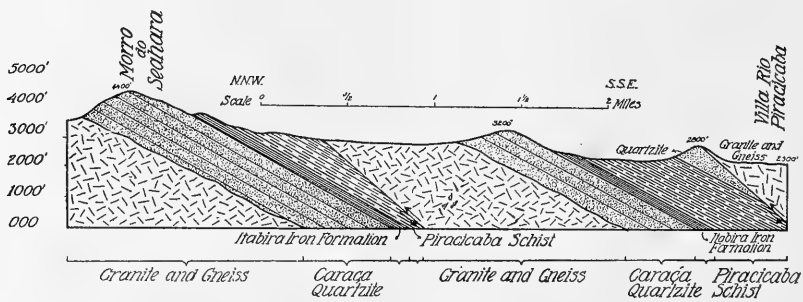

FIG. 7.—Slice-faulted section between Villa Rio Piracicaba and Morro do Seahara

The type of faulting can perhaps best be described as slice faulting (Fig. 7). The strata, instead of yielding to the stresses by wrinkling and folding, were sliced off in a succession of inclined blocks nearly all dipping toward the east or southeast, and the crustal shortening accomplished by movements between the inclined slices. In the northern portion of the Minas Geraes iron-ore district, the slice faulting has caused an alternation of the crystalline complex with the sedimentary series. Nearly everywhere the basement complex is the surface formation on the upthrust side. On account of the great erosion which the region has suffered since the period of deformation, the sedimentary formations have persisted only on the downthrown side of each fault. In the western part of the district also, faulting would seem to dominate folding.

Some of these fault lines are apparently of great length. A great fault forms the eastern front of the Serra do Espinhaço, or Backbone Ridge, at Cocaes and from there runs northward for many miles. Then, as we go farther north, there appear in front of it to the east successively new faults arranged *en échelon*. Each of these, in turn, commences a new front range slightly to the east of the previous range. The old ranges continue northward as successive ridges behind the front. This system of faulted ranges runs northward into the state of Bahia.[1] The displacement along the major fault planes cannot be determined, inasmuch as the sedimentary formations have completely disappeared from the upthrust side, so that the crystalline complex alone remains. It is probably great. In the minor slice faults the displacement in some cases is as low as 300 meters. In general the faulting in central Minas Geraes shows a marked similarity to that in the Appalachian region of the United States and that in the Canadian Rockies of Alberta.

These deformative movements were very general, apparently involving most of the Atlantic border of Brazil south of Cape St. Roque, and possibly also Uruguay and portions of the province of Buenos Aires in Argentina. The mountains formed at this time would seem, in a way, to be the Atlantic counterpart of the Andean ranges on the Pacific border. South America was squeezed from the Atlantic first and from the Pacific later.

How closely the period of deformation followed the sedimentation is as yet unknown. Minas Geraes affords little evidence, but in several of the neighboring states, Matto Grosso, São Paulo, and Parana, strata of Devonian age, but slightly disturbed, are found resting unconformably upon the wrinkled strata which participated in these orographic movements.[2] The mountain-building movements were therefore accomplished before the Devonian. That the deformation did not immediately precede the Devonian is evident from the amount of erosion suffered by the mountains

[1] O. A. Derby, *op. cit.*

[2] J. E. Branner, *Geologia Elementar*, pp. 230–31; O. A. Derby, "Geologia da região diamantifera da Provincia do Parana, no Brazil," *Archivos do Museu Naçional de Rio de Janeiro*, III (1878), 89–96.

before the incursion of the early Devonian sea. In São Paulo and Parana, the basal beds of the Devonian (which have been assigned to Lower Devonian) rest unconformably on a westward dipping, now slightly warped surface of these old deformed rocks in such a manner as to show that the sea crept in over a region of little or no relief.[1] An interval of much erosion is therefore implied.

The location and trend of these mountain ranges is worthy of note for their bearing upon the major problem of earth movements and deformation. It is no new fact that the ranges of eastern Brazil are rather strikingly parallel to the border of the continent, and not far back from the coast. In Minas Geraes the repeated overthrust faulting has nearly all come from the southeast. The strata of the faulted slices nearly always dip toward the Atlantic. The picture of the Atlantic Ocean segment crowding the eastern side of the South American continent, squeezing it and forcing it upward like a triangular wedge and wrinkling and slicing the edge comes very appealingly to mind. At a later time the Pacific Ocean segment seems to have done the same on that side.

PALEOZOIC—EARLY MESOZOIC EROSION PERIOD

After the great period of mountain-building we have evidence only of a long period of erosion whose final result was to develop an approximate base-level which is, at present, only preserved on the crest of the Serra do Espinhaço. The amount of material removed from the upthrust blocks was great, suggesting that the interval between the end of the deformation and the completion of the base-level was a long one. The process of base-leveling was probably not rapid, since the hard Caraça quartzite is a very resistant formation and would have been reduced slowly. That the process progressed nearly to completion is evidenced by the character of the crest of the Serra do Caraça which, even today, preserves in places a well-developed plain (Fig. 8). The Serra do Caraça being composed of more resistant rock than the surround ng region, this quartzite area would be the last portion of the region to be brought to base-level. As the upturned Caraça strata were

[1] J. B. Woodworth, "Geological Expedition to Brazil and Chile, 1908–9," *Bull. Mus. of Compar. Zool. Harvard College*, LVI, 42.

truncated, it is safe to assume that all the rest of the region was close to base-level.

MESOZOIC—EARLY TERTIARY DEPOSITION

Diamantina conglomerate.—On the truncated, even crest of the Serra do Espinhaço, high-level gravels, now cemented into a conglomerate, are found at various points along the range, particularly

FIG. 8.—Peneplained crest of the Serra do Caraça, near Alegria. The inclined quartzite strata have been trunkated and the old erosion surface still persists over extensive areas on the top of the Backbone Range.

in the vicinity of Diamantina. The top of the Backbone Range in that region is very broad, partaking perhaps rather more of the nature of a narrow plateau belt than of a mountain range. The broad open tracts of comparatively level upland are so distinctive as to have been termed *chapadas* (table-lands) by the natives. While considerably dissected by subsequent erosion, the broad open country upon the top of the Backbone Range is the modified surface of the old peneplain (Fig. 9). At various points on these *chapadas*, or upland plains, especially in hollows or former river

channels in the old surface, are found deposits of residual clay and gravel, more or less imperfectly cemented. This material is all thoroughly weathered and of a residual nature. The pebbles in the conglomerate are mostly of quartzite, but with a sprinkling of pebbles of iron formation, schist, quartz, basic igneous rock, and diamonds. With the conglomerate is associated a considerable amount of sand and of clean white to pinkish clay, mostly kaolin, with some bauxite. This material for the most part represents

FIG. 9.—A *chapada*, or table-land, forming the summit of the Backbone Range near Diamantina. The diamond-bearing conglomerates are accumulations of residual gravels collected in depressions or old stream channels on the peneplained surface.

the final products of the rock weathering, and its presence perched upon the smooth crest of the Serra do Espinhaço above the rest of the region would suggest that it is to be connected with the base-level stage in the history of the region. This is the celebrated diamond-bearing conglomerate which has brought fame and wealth to the district, and which would be interpreted in physiographic studies as the high-level gravels which mark the peneplain stage. Subsequently, during later erosion cycles, some of the conglomerate

material has been washed down on to lower benches and terraces along the courses of some of the streams which dissect the plateau, and today diamond mining operations are also carried on at these lower levels.

LATER EROSION AND DEPOSITION PERIODS

General uplift.—As to the age of the peneplain nothing definite is known, but it would seem most likely to belong to the late Mesozoic or early Tertiary. At some time subsequent to the formation of the plain, the region was uplifted as a whole with but little change in the attitude of the strata. Rejuvenated erosion then commenced to carve the present mountains out of the uplifted plateau. The resistant quartzites remained as ridges, while the softer schists and readily disintegrated granites and gneisses were more rapidly worn down and removed. In this tropical country the granites and gneisses appear to be the least durable of the formations, owing to the rapidity with which they crumble to pieces following the weathering of their ferromagnesian constituents. The micas, amphiboles, and pyroxenes are readily attacked by humic acids and by the carbon dioxide of the atmosphere. Weathering of these constituents allows the crystalline mass to crumble into arkose which is removed by the streams. The result is that in Minas Geraes the areas of granite, because of the feeble resistance of this kind of rock to the accelerated chemical action characteristic of weathering in the tropics, have become the lowlands. Some prominent granite ridges occur, but on examination it is generally found that they are composed of light-colored granite in which the ferromagnesian constituents play a subordinate part. The quartzite formations in general owe their resistance to erosion mainly to the fact that they contain a very small percentage of readily decomposable minerals.

Tertiary canga plains.—The erosive processes which have carved the mountains of central Minas Geraes from the uplifted peneplain did not go on uninterruptedly. There were times of checked erosion, locally at least, followed by rejuvenation. These lesser cycles are shown by various local plains and remnants of plains in different parts of the region. One of the most notable of these is

the high-level canga plain in the locality known as Gandarella, 40 kilometers northwest of Ouro Preto.

Canga (from Itapanhoacanga, a village in central Minas) is a peculiar ferruginous conglomerate consisting chiefly of fragments of iron formation cemented with a nearly pure iron-oxide cement. It is formed by a surface disintegration of the iron formation which consists largely in the removal of silica, and solution and redeposition of iron oxide. In some places the canga has been deposited almost *in situ*, while elsewhere it has been transported for considerable distances and deposited, either on areas of iron formation, or on adjacent areas of other rocks.

Such canga deposits have probably been forming ever since the iron-formation beds were exposed to surface erosion, and are being abundantly formed at the present time. As a consequence they are found at many different elevations. A long period of quiescence, like that in which the land was worn low following the pre-Devonian deformation, was therefore probably accompanied by abundant canga deposition which formed extensive canga plains on, and adjacent to, the belts of iron formation. But subsequent erosion has been so extensive that today few remnants of canga formed during the great base-leveling period remain.

The Gandarella canga plateau, already mentioned, is the most conspicuous of the high-level canga plains and is situated on the summit of the divide separating the waters of the upper Santa Barbara valley from those of the upper valley of the Rio das Velhas. This canga plateau has an approximate elevation of 1,450 meters, a little lower than the Serra do Caraça. Whether it is to be regarded as a portion of the Caraça peneplain, now at a lower level than the rest of the plain, and so more or less contemporaneous with the Diamantina conglomerate, or whether it is to be taken as the work of a later cycle of erosion, and to be correlated with evidences of planation at a similar elevation in Alegria and near Antonio Pereira, may perhaps best be left open for the present. Deep valleys have been cut in this remnant of the canga plateau along both sides of the divide. Along the bottom of one or two of these at Gandarella, there occur irregular deposits of clay and sand containing lignite and fossil leaves. On the basis of the plant

remains, these deposits have been referred to the middle or late Tertiary—probably Miocene or Pliocene.[1] They are distinctly younger than the canga plateau which is therefore certainly older than the Pliocene, and probably older than the Miocene.

Similar deposits of Miocene or Pliocene clay, sand, and lignite occur near the village of Fonseca,[2] east of an extensive canga plain of later age, which occurs along the eastern front of the Serra do

Fig. 10.—A plain of canga formation, between Santa Rita Durão and Agua Quente.

Caraça. This younger plain is very conspicuous over the areas of softer formations east of the Serra do Caraça. The canga plain abutting the east foot of the Serra do Caraça between the village of Agua Quente and Santa Rita Durão (Fig. 10) is a particularly well-preserved portion of it. The softer schists, gneisses, and granites east of the present Serra do Caraça had, with the exception of some of the more resistant lenses in the iron formation, been

[1] H. Gorceix, "Bacias terciarias de agua doce nos arredores de Ouro Preto," *Ann. da Escola de Minas de Ouro Preto*, III, 95–114; Rio de Janeiro, 1884; Joaquim Candido da Costa Sena, *Ann. da Escola de Minas*, N. 10 (1908), p. 17.

[2] H. Gorceix, *op. cit.*

reduced to a common level. Upon this plain, material washed down from the iron-formation monadnocks developed a covering sheet of canga. The canga formation, by obliterating some of the remaining irregularities in the erosion surface, has produced a remarkably level plain, much of which still remains. This canga plain is now at an average altitude of about 900 meters above the sea. Judging from the occurrences at Fonseca it appears to be of later age than the lignite deposits, which have been referred to the Miocene or Pliocene, and would seem to belong to the late Tertiary. Over considerable areas it has scarcely been trenched at all.

Conglomerate formations.—Benches, and other remnants, corresponding to the canga plains in the neighborhood of the iron formation, are also to be seen in the Diamantina region. These remnants occur in the neighborhood of the streams which have worked back and partially dissected the plateau-like Backbone Range. As the formation of the range is quartzite, the gravels developed on these plains are quartzose, though possibly to be correlated with the iron-formation conglomerate, or canga, of the iron-ore district. As these gravels and conglomerates are composed of concentrated residual material derived from diamond-bearing formations, they have been extensively worked for diamonds.

PRESENT STATUS OF EROSION

The formation of these several Tertiary plains was terminated supposedly by uplifts which accelerated stream action and started their dissection. Portions of the canga plain near Agua Quente have suffered considerable erosion since its formation. Toward the northeast, in the region between the Serra do Caraça and the Piracicaba iron-ore district, the plain has been much carved. This area being for the most part farther from the hills of iron formation, the thick capping of tough canga which has preserved the plain intact at Santa Rita Durão and Agua Quente was much less developed. As a result, erosion progressed more readily. The easily weathered underlying granites and gneisses which almost exclusively make up this region have been cut into a state of mature erosion, though the level of the former peneplain is well preserved in the tops of these low hills.

As no attempt was made to correlate erosion cycles in the different parts of the district, little can be said of these upward movements of the land surface. One such uplift of the plateau-forming type which affected the coastal region of Brazil in this latitude was that which has given rise to the Serra do Mar, or Coast Range. The effects of this movement were undoubtedly felt by the drainage systems farther inland in Minas Geraes. The Serra do Mar presents a very abrupt face toward the sea. At many points it is really a plateau scarp which has the appearance of a mountain range when viewed from the coast. Judging from the abruptness of this plateau scarp, the upward movement which produced it must have occurred late in the Tertiary, or possibly even early in the Pleistocene.

Present topography.—The present topography of central Minas Geraes has resulted, to a very large extent, from differential erosion of rock formations which offer varying resistance to the processes of degradation. The mountains and ridges, valleys and plains, have been carved out of a region which has been folded and faulted and afterward reduced to base-level. The location of the present topographic features has been determined by the underlying rock structure. Wherever, because of faulting or folding, the heavy quartzite formations appeared at the surface of the plain out of which the present topography was sculptured, they resisted erosion far more than the other rocks, and now remain as mountain ridges. The iron formation, at most points somewhat less resistant than the quartzite, but throughout most of its extent more lasting than adjacent schists, stands up as a chain of foothills in approximately parallel arrangement to the higher quartzite ridges. The schists, by yielding more readily to erosion, have developed comparative lowlands in the midst of which isolated ridges appear wherever a lens of quartzite, or hard iron formation, occurs.

But the conspicuously weak formation is the basement complex. The granites and gneisses are especially susceptible to the chemical disintegration which is induced by the tropical conditions. Their ferromagnesian constituents are so readily attacked by the carbon dioxide of the atmosphere and the humic acids from the rank vegetation, aided by the hot, humid climate, that the rock rapidly

crumbles to pieces and the crystalline areas become lowlands. But on account of comparatively recent uplifts, the granite areas of central Minas Geraes are not yet reduced to plain conditions, though they represent a state of advanced erosion. The hills are mostly small and very numerous. They produce the peculiar knobby, hummocky topography which characterizes so many areas of crys-

FIG. 11.—View toward the Serra do Caraça from the Morro do Seahara showing the topography resulting from the unequal resistance of the formations; quartzite ranges in the foreground and in the distance; Morro Agudo, a pyramidal peak of iron ore, to the left of the center of the picture, and the billowy lowlands of granite beyond.

talline rock within the tropics, and which has been aptly likened by Eschwege to the billows of a rolling sea (Fig. 11). Wherever erosion has become sluggish in these lowland areas, the rapid disintegration of the crystalline rock has formed a very thick covering of mantle rock which effectively conceals the solid rock from view.

Reviews

An Introduction to the Geology of New South Wales. By C. A. Süssmilch. Sydney: Angus & Robertson Ltd., 1914. Pp. 269, figs. 92, tables 6, with folding colored map.

The present volume is a revision and enlargement of the first edition which appeared in 1911 and which was then welcomed as treating in a concise and judicious manner the more important features of the geology of New South Wales. After devoting one chapter to physiography, detailed historical geology is taken up, period by period. One of the most commendable features of the treatment, and one which will be especially appreciated by geologists from other countries, is the summary at the close of each chapter in which the essential and significant features of the period are given in their proper relations. From this the reader gets the correlations and conclusions of the author in a most convenient and serviceable form for use.

In New South Wales, orogenic movements are recorded in the pre-Cambrian and Paleozoic eras only. The most important of these movements appear to have taken place at the close of the pre-Cambrian, at the close of the Ordovician, at the close of the Devonian, and, in the northeastern portion of the state, also at the close of the Carboniferous and of the Permo-Carboniferous periods. But it is worthy of note that throughout the Mesozoic and Cenozoic eras, the crustal movements, not only of New South Wales, but also of Australia as a whole, appear to have been almost entirely of the epeirogenic type.

Bearing upon the existence of the hypothetical Gondwana land, in the opinion of the reviewer, is Süssmilch's statement that from New South Wales, in common with the rest of Australia, the extraordinary group of terrestrial and flying reptiles which dominated the Mesozoic land life of Europe and North America is conspicuously absent, the only vertebrates known to have lived during this era being fish, amphibia, and a few marine reptiles. If Australia had been connected by land with Asia, and especially with South Africa, where terrestrial vertebrates were so remarkably deployed and developed early in the Mesozoic, these animals should have invaded Australia also. The higher terrestrial reptiles seem to have been as completely cut off from Australia during the Mesozoic era as were the placental mammals in the Cenozoic.

R. T. C.

Oklahoma Geological Survey Bulletin 22. Part I, "Director's Biennial Report"; Part II, "Mineral Resources of Oklahoma and Statistics of Production from 1901 to 1914." By C. W. SHANNON.

Bulletin 13. Volcanic Dust in Oklahoma. By FRANK BUTTRAM.

The great progress in the development of the mineral resources in the state of Oklahoma in the last ten years is made clear by the following statistics:

VALUE OF MINERAL PRODUCTS OF OKLAHOMA

	1903	1913
Coal	$6,386,463	$8,542,748
Petroleum	142,402	59,581,948
Natural gas	1,000	7,436,389
Asphalt	28,150	91,416
Granite	9,030	30,678
Sandstone	6,500	1,010
Limestone	56,140	246,912
Line	4,800	12,160
Cement	....	1,258,676
Gypsum	234,621	330,416
Clay products	534,977	573,371
Sand and gravel	....	39,457
Lead	....	548,064
Zinc	....	1,306,368
Salt	2,070	259
Mineral water	....	26,231
Other products	....	5,527
Total	$7,406,153	$80,031,630

The report is accompanied by maps showing the distribution of the mineral resources in the state, as well as a map showing its physiographic divisions.

Volcanic dust has been found at widely separated localities in the state. In one place it has a thickness of 9 feet. The dust rests upon formations as young as Pliocene, and is believed to have come from sources 600 miles or more away. R. D. S.

"A Contribution to the Optical Study of the Amphiboles." By W. E. FORD. *Am. Jour. Sci.*, XXXVII, 179–93, figs. 11.

A study with a view to correlating the optical properties of analyzed specimens with their chemical composition. Extinction angles were found

to be nearly identical in specimens of notably different composition. In general the extinction angle seems to decrease with increasing iron, but the author points out that the composition may vary in a complex way and no very sure conclusions can be drawn.

A. D. B.

Ninth Report of the Director of the Science Division. By JOHN M. CLARKE. New York State Museum Bull. 164. 1913. Pp. 241, pls. 46.

This publication includes the 66th Report of the State Museum, the 32d Report of the State Geologist, and the Report of the State Paleontologist for 1912. Although unaccompanied by descriptions, there are 36 plates illustrating the Devonian fossils of Brazil and of the Falkland Islands. A full description of the fauna is to be published by the director of the Geological Service of Brazil.

The object in printing these plates is to show the distinctive characters of the southern Devonian fauna. Although fundamentally related, it is apparent that the genera of the northern and southern American habitats developed in separate basins with restricted communication. From the fossils there is indication of a continuous strand line from South America through the Falkland Islands to South Africa.

T. T. Q.

Stratigraphy and Paleontology of the Alexandrian Series in Illinois and Missouri. Part I. By T. E. SAVAGE. Extract from Bull. 23, Illinois State Geol. Survey. 1913. Pp. 124, pls. 7.

This part of the subject treats only the Girardeau and Edgewood limestones. Part II on the Essex and the Sexton limestones is in preparation. The report discusses the stratigraphic and paleontological relations of the series. A large part is a description of the fossils.

Alexandrian series is the name proposed for the early Silurian strata of Illinois and eastern Missouri which occupy a position above the top of the generally accepted Richmond and below the Brassfield (Ohio Clinton) limestone. The series consists of a number of closely related formations; these record a succession of oscillatory northward sea advances separated by breaks in sedimentation due to temporary withdrawals of the sea. The Girardeau was the earliest and the least extensive invasion; the cycle closed with the most widespread, the Sexton Creek submergence.

T. T. Q.

Geological Report on Arenac County. By W. M. GREGORY. Michigan Geol. and Biol. Surv. Publ. 11, Geol. Series 8. 1911. Pp. 148, pls. 6, figs. 18, map 1.

The county is underlain by Mississippian and Pennsylvanian formations which dip slightly to the south. The economic resources are very slight. Limestone, gypsum, and clay are of local usefulness.

T. T. Q.

Annotated Bibliography of Iowa Geology and Mining. By CHARLES KEYES. Iowa Geol. Surv., Vol. XXII. 1913. Pp. 908.

Most of the first 150 pages of the bibliography are given to historical material. The bibliography is arranged alphabetically under the names of authors and subjects.

T. T. Q.

Prince George's County. By WILLIAM BULLOCK CLARK. Maryland Geol. Surv. 1911. Pp. 251, pls. 13, figs. 3. Accompanied by *Prince George's County Atlas,* 2 maps.

The sixth of a series of reports dealing with the physical features of the several counties of Maryland. A full discussion of the stratigraphic geology of the county accompanies a description of the physiography, mineral resources, soils, forests, climate, and hydrography.

The geologic formations represented in the county range from Archean to very recent. After the Archean, no formations are represented below the Potomac group of the Comanchean. Later formations represent the Cretaceous, the Eocene (Pamunkey), the Miocene (Chesapeake), and later periods.

T. T. Q.

The Manhattan Schist of Southeastern New York State and Its Associated Igneous Rocks. By CHARLES REINHARD FETTKE. A dissertation (Columbia). Annals N.Y. Acad. Sci., XXIII, April 30, 1914, pp. 193–260, Plates VIII–XV.

The erosion of northeast-southwest trending anticlines and synclines has exposed the Manhattan schist in a series of roughly parallel belts south of the Highlands of the Hudson and east of the Hudson River. The Manhattan is a quartz-mica-feldspar schist and the young-

est bedrock formation of the region. It is believed to be a greatly metamorphosed series of "argillaceous and sandy shales, argillaceous sandstones, and arkoses which represent a thickness of several thousand feet." The associated igneous rocks are, in the order of decreasing age, basic sheets and flows, granitic intrusives (batholiths with radiating dikes of granite and pegmatite), basic intrusives, and granite and pegmatite intrusives, a basic dike. The Manhattan schist apparently overlies the Inwood limestone conformably. The Inwood-Manhattan series is thought by Merrill, Dana, Mather, and others to be equivalent to the Cambro-Ordovician Poughquag-Wappinger-Hudson River series; it is believed by Berkey to be pre-Cambrian.

V. O. T.

The Constitution of the Natural Silicates. By F. W. Clarke. U.S. Geol. Surv. Bull. No. 588. Pp. 128.

In the opening chapter the author outlines some of the bases upon which structural formulae may lie, but in the remainder of the bulletin the structures are worked out by simply matching valences in such a way as to agree with the empirical formulae. Until our methods of synthesis are better worked out, and the decomposition of silicates is better understood, it is difficult to justify the speculative structures advanced, as they do not rest on a foundation of experimental study, but rather on the more mathematical concept of valence and chemical combination.

A. D. B.

Our Mineral Reserves. By G. O. Smith. U.S. Geol. Surv. Bull. No. 599. Pp. 48.

In response to a demand for information as to sources of various mineral products the director of the Geological Survey has prepared this bulletin, which deals with the general situation, and briefly summarizes the condition of the industries producing some twenty-odd products.

A. D. B.

The Darwin Silver-Lead Mining District, California. By Adolph Knopf. U.S. Geol. Surv. Bull. No. 580-A, pp. 1–18. Figs. 3.

Some of the ore bodies in this region are of contact metamorphic origin, and some are transitional; but most are fissure veins. For the most part, the ore bodies are found in the "lime-silicate" rocks which are metamorphosed sedimentary rocks. The deposits are of interest in that they show the transition between contact metamorphic deposits and fissure veins.

A. D. B.

Notes on the Unaweep Copper District, Colorado. By B. S. Butler. U.S. Geol. Surv. Bull. No. 580-B, pp. 19–23.

Notes of a two-day reconnaissance. The ores occur in fissures cutting both igneous and sedimentary rocks. The district is practically idle, and has never advanced beyond the prospecting stage.

A. D. B.

Some Cerusite Deposits in Custer County, Colorado. By J. Fred Hunter. U.S. Geol. Surv. Bull. No. 580-C, pp. 24–37. Figs. 5.

Describes the deposits of cerusite 12 miles northeast of Silver Cliff. The ores are in a crushed fault zone. The mines have been idle for more than ten years. The two days spent in the field by the author preclude the possibility of any exact statements regarding either the general geology or the ore deposits.

A. D. B.

The Grand Gulch Mining Region, Mohave County, Arizona. By James M. Hill. U.S. Geol. Surv. Bull. No. 580-D, pp. 38–58. Figs. 10.

"The ore bodies occur around the sides of a plug-like mass of rock, which is sedimentary in origin but is entirely unlike the rocks that enclose it." The ore bodies, carrying copper, are more or less lens-shaped replacements of the rocks in which they are found. The author shows interesting drawings of the relation existing between chalcopyrite, bornite chaleocite, and malachite, as successive steps in the alteration of the primary sulphide.

A. D. B.

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THE GEOLOGY OF CENTRAL MINAS GERAES, BRAZIL

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PART II

PRINCIPAL MINERAL DEPOSITS

IRON ORES

On the basis of their physical properties, which to some degree separates them also according to chemical composition, the iron ores[1] of Minas Geraes may be classified as follows.

[1] H. K. Scott, "The Iron Ores of Brazil," *Jour. Iron and Steel Institute*, LXI, (1902), 248; O. A. Derby, "The Iron Ores of Brazil," *Iron Ore Resources of the World*, p. 313, Stockholm, 1910; C. K. Leith and E. C. Harder, "The Hematite Ores of Brazil and a Comparison with the Hematite Ores of Lake Superior," *Econ. Geol.*, VI, (1911) 670–86; E. C. Harder, "The 'Itabirite' Iron Ores of Brazil," *Econ. Geol.*, IX, (1914) 101–11; E. C. Harder, "The Iron Industry in Brazil," *Bull. Am. Inst. Mg. Engrs.*, October, 1914.

- Original or Bedded Ores
  - Hard massive ore
  - Soft powdery ore
  - Laminated or thin-bedded ore
- Concentration Ores
  - Canga
  - Stream sand and gravel ores
  - Rubble ore
  - Enriched itabirite
  - Leached carbonate

Of these the bedded ores, especially the hard massive ores, are by far the most important. Of the concentration ores the canga is the principal type.

Bedded ores.—Bedded iron ores are found in many places in central Minas Geraes, but the most important deposits occur in the region surrounding the headwaters of Rio Piracicaba, Rio Carmo, Rio das Velhas, Rio Santo Antonio, and Rio Paraopeba, together with their various branches and tributaries. This is the region in which the Itabira iron formation has survived the extensive denudation which has swept away the post-Archean formations from much of eastern Brazil, and where it still remains as a thick formation over considerable areas. The ore bodies occur at various points along the belts of iron formation already described, both in the Itabira iron formation proper, and in the iron-formation lenses in the Piracicaba schist. They are not irregularly scattered throughout the iron-formation areas, but are rather grouped in certain portions with barren areas between. Thus groups of iron-ore deposits occur north of Ouro Preto on the east slope of the Serra do Caraça;[1] on the Rio Piracicaba west and northwest of the town of Villa Rio Piracicaba; on the prominent peaks at Itabira de Matto Dentro; along the ridge running from Sabara northeastward to Caethé and southwestward beyond Bello Horizonte; at various points in the area west of Itabira do Campo, and in the region west of Burnier and north of Gongonhas do Campo. Smaller, less important groups occur elsewhere in the district.

The bedded ores occur as beds or lenses in the iron formation. Some are thin and continuous for long distances, while others are

[1] See map on p. 347, Part I of this article.

thick and of small diameter. Of the three classes of bedded ores, hard massive ore and the peculiar soft powdery ore are the most widespread, the latter probably being the more common of the two. Laminated ore is less widespread, but generally occurs in larger deposits.

The hard ore of central Minas is of several varieties, the most common of which is a slightly specular blue hematite resembling

FIG. 12.—The Peak of Cauê rising above the town of Itabira de Matto Dentro. A large body of hard ore.

some of the specular hematite ores on the Marquette range in Michigan. Such ores occur in many places in the district. Locally there are deposits consisting of coarsely specular hematite with scattered crystals of martite, and with these are frequently associated ores composed largely of coarse granular masses of martite, or magnetite not yet completely altered to hematite. Such phases are due probably to greater metamorphism and are most common in the eastern part of the district. In the central and western part of the district there occur a few deposits of a tough, fine-grained, almost amorphous hematite, or hematite intermixed with some limonite.

Some hard ore deposits, or parts of these deposits, are so massive that the bedding is barely distinguishable while other portions may be distinctly bedded. The material is generally well consolidated, and some phases, especially the amorphous variety, are of extraordinary hardness and toughness.

Of the different varieties of hard ore the finely specular type is the highest grade. It generally averages between 69 and 70 per cent in metallic iron and rarely has more than 0.025 per cent phosphorus and frequently runs as low as 0.003 per cent phosphorus. Other impurities are practically absent.

The soft, powdery ore deposits consist largely of unconsolidated material of great fineness, most of it so fine-grained that it will easily pass through a 100-mesh screen. Although fine, the material is almost entirely crystalline, quite different from the amorphous powder which occurs with much of the soft ore on the Mesabi Range in Minnesota. Some of the powdery ore is slightly consolidated when moist, but readily crumbles when dried. Most of it, however, is soft in the moist state like fine moist sand.

The powdery ore is dark blue in color like the hard ore and resembles it in composition, though in general perhaps of slightly lower grade as it is more frequently intermixed with a sprinkling of quartz sand. It averages between 67 and 69.5 per cent metallic iron and contains up to 0.05 per cent of phosphorus, though it may run as low as 0.004 per cent of phosphorus. When the amount of quartz sand increases in the soft ore the metallic iron content decreases until the arbitrary figure of 50 per cent metallic iron is reached, when the material is classified as itabirite. Some of the soft ores, just as in the case of some of the hard ores, are almost without evidence of bedding, while others are distinctly bedded and even finely laminated. Where quartz sand is intermixed with the soft ore the bedding is well marked.

Hard ore and soft ore are generally more or less associated with each other. In general it may be said that massive specular ore and soft, powdery ore are hard and soft phases of the same material (Fig. 13). A deposit composed principally of hard ore may have lenses of soft ore or irregular masses of soft ore scattered through it, or portions of it may be composed of a mixture of hard

ore fragments and soft ore, the latter acting as a matrix. On the other hand, deposits consisting largely of soft ore may have hard ore lenses interbedded with the soft ore.

The soft, powdery ore is known among the natives of the iron region as "jacutinga," and since this term has also crept into the literature its meaning should be clearly defined and its misuse avoided. The term "jacutinga" was applied by Hussak[1] to

FIG. 13.—The Peak of São Luis near the village of Agua Quente. In front of the peak is a plain surfaced by canga. The steeply inclined beds of iron formation which appear in the peak continue under the canga covering of the plain. In the peak they consist partly of itabirite and partly of hard ore; under the plain the same beds consist of soft, powdery ore and soft itabirite. On the extreme left is the Caraça quartzite range.

certain gold-bearing portions of the iron formation. The jacutinga thus designated occurs in the form of shoots and irregular masses inclosed in itabirite from which it differs in a minor way only. It has a variable composition, and its principal characteristic is a fine

[1] E. Hussak, "O Palladio e a Platina no Brazil," *Annaes da Escola de Minas de Ouro Preto*, N. 8 (1906), p. 96, tr. by Miguel A. R. Lisboa, and Manoel A. R. Lisboa.

micaceous texture. It consists mainly of soft hematite in which quartz occurs in varying abundance together with talc, kaolin, mica, and earthy pyrolusite and frequently tourmaline. Presumably the term as used by Hussak includes those portions of the iron formation which have been affected by gold-bearing solutions, and it seems best to limit its use in the future to this phase and to avoid its use in the iron-ore terminology.

The bedded ores of the third class, the laminated, are distinct from those of the other classes and more closely resemble the itabirite in texture. These ores might be termed itabirite with little or no quartz sand. They occur in large lens-like deposits in the iron formation, as do the hard ores and soft ores, but the boundary between one of these lenses and the itabirite of the iron formation which incloses it is less well defined and more irregular. While the hard ore and soft ore deposits generally have a sharp contact with the inclosing itabirite, the laminated ore deposits generally have a zone of gradation with the itabirite and also have interbedded masses of itabirite within them. The separation of itabirite from laminated ore is frequently quite arbitrary, being controlled by the percentage of metallic iron.

Lenses of hard ore or soft powdery ore occur interbedded within laminated ore deposits in the same way that they occur in the itabirite portions of the iron formation. In such cases the contact with the hard ore is well defined, but that with the soft ore less so.

Laminated ore is a very thinly bedded, porous ore which is quite friable, breaking into thin plates along bedding planes, owing to the fact that the porosity is concentrated along these planes. The ore is red or blue hematite, much of it in the amorphous form, with a considerable percentage of limonite. The different varieties of laminated ore are generally found interbedded or interlaminated. Certain beds may be red, others blue, and still others yellow, the differences being the result of hydration and slight admixtures of clay and other impurities. Frequently siliceous beds of different thicknesses will be found interbedded with pure ore.

The laminated ore is not so high in grade as the hard ore and soft ore, for the metallic iron content of its richer portions varies

between 63 and 67 per cent, while the hydrated portions frequently contain as high as 0.3 per cent phosphorus and rarely less than 0.05 per cent phosphorus. It is mainly a non-Bessemer ore. With increasing quartz sand the content of metallic iron decreases until the material becomes itabirite. Non-hydrated, laminated ore resembles soft, powdery ore in composition.

The phosphorus content of the surface ores is always higher than that of the ore underground. Where surface ore contains 0.12 to 0.18 per cent phosphorus, underground ore may contain only 0.04 to 0.08 per cent. This is due to the concentration of phosphorus during the process of weathering. This same process also concentrates the iron in the uppermost foot or two, hardening it and decreasing the pore space.

The size of the different ore deposits varies from mere seams inclosed in itabirite to beds several kilometers in length and several hundred meters in thickness. The largest known deposits are of laminated ore, but there are also hard ore deposits of large size. The largest known laminated ore deposit in the district is that of Alegria, where continuous outcrops occur over a distance of more than 4 kilometers along the strike of the beds and for fully 1,700 meters across the beds. With an average dip of 35° the thickness would be at least 1,200 meters.[1] This great thickness, however, is not entirely of pure ore, but will possibly be found to include numerous minor beds of itabirite now concealed at the surface. The largest body of hard ore in the district, which, however, has a considerable admixture of soft ore and of hard ore fragments intermixed with soft ore, is that of Periquito near Itabira de Matto Dentro. It has a length of 1 kilometer and a maximum thickness at right angles to the beds of 250 meters. Other large deposits of hard ore are known in which the proportion of soft ore is very small. Such are the peak of Caué (Fig. 12), the deposits on Serra de Conceiçao and Serra do Esmeril near Itabira de Matto Dentro, Itabira Peak near Itabira do Campo, and Morro Agudo Peak near Villa Rio Piracicaba. Deposits consisting largely of pure soft powdery ore are as a rule much smaller than hard ore deposits but more numerous.

[1] See Fig. 4 on p. 354, Part I of this article.

Concentration ores.—Concentration ores occur in the neighborhood of other ore deposits or of iron-formation beds by the disintegration of which they are formed. They are in the form of surface deposits and do not go to a great depth. Of such deposits the canga ores are the most widespread, occurring as surface blankets of great horizontal extent over a large portion of the areas of iron formation, as well as over areas of other rocks bordering the iron-formation

FIG. 14.—The Fazenda de Alegria. The prominent mountain is the Serra do Caraça. The lower range of hills in the middle distance are the hills of laminated iron ore. The ranch house is built upon canga-covered Piracicaba schist.

belts. They are composed of material resulting from the disintegration, not only of the iron ore portions of the iron formation, but from the itabirite phase as well.

Canga consists of different sized fragments of iron ore mixed in places with fragments of itabirite, the whole cemented together by iron oxide. The incorporated fragments vary in size from sand grains to bowlders weighing perhaps several tons. Some of them consist of pure massive ore, others of laminated ore or itabirite. The fragments have been mechanically concentrated and have

been cemented together by iron oxide deposited from solution, both fragments and cementing material being derived from the same source—the iron formation. The cementing material is largely hydrated hematite and limonite, red or yellow in color, giving different portions of canga deposits colors varying from light yellow to dark red. The fragments decrease in size and abundance as the distance from the source increases, so that in one place a canga deposit may consist largely of cemented fragments and elsewhere of finely textured chemically deposited iron oxide. The latter is most common over areas outside the iron formation or over soft portions of the iron-formation belts where erosion yields no coarse fragments. In the latter localities, however, much fine fragmental material is intermixed. In places at a distance from the iron-formation areas the canga is very low grade, containing much clay and quartz sand, the iron oxide present being principally limonitic cement.

The ordinary canga of which the great portion of the canga deposits are composed averages between 60 and 65 per cent metallic iron and up to 0.3 per cent phosphorus, rarely containing less than 0.1 per cent phosphorus. This concentration of phosphorus is due to the same cause as the concentration of the phosphorus in the surface laminated ore. The outlying portions of canga deposits at a distance from the iron-formation belt are so low in metallic iron that they do not constitute an iron ore.

A canga blanket may vary in thickness in different portions from a few centimeters to 20 meters or more. It is generally thickest on the lower slopes and at the base of iron-formation ridges, where the conditions for accumulation are best, and decreases in thickness both up the slope and toward the valley. A single continuous blanket of canga may cover several square kilometers.

Stream deposits of iron oxide in the form of sand, pebbles, and larger fragments occur in greater or less purity along most of the streams rising in or flowing through areas of iron formation. The material is distributed over the width of the valley floors and in places occurs in terraces several meters above the present valley bottom. The stream deposits do not constitute a very high-grade ore, always being mixed with more or less quartz sand and other

foreign material. They are of little or no importance as an ore reserve.

Rubble ore deposits occur on the hill slopes below outcrops of hard ore. They are simply talus accumulations consisting of high-grade iron ore, and frequently are large enough to be of some importance for mining purposes. The fragments composing them may vary in size from pebbles to fragments of several tons. In

FIG. 15.—A cliff of canga on the side of the "Grotto" in Esmeril near Itabira de Matto Dentro. The canga blanket here reaches a thickness of about 15 meters.

many places, however, one finds rubble ore deposits with fragments remarkably uniform in size. Rubble ores have practically the same composition as the hard ore deposits with which they are associated.

Small deposits of enriched itabirite occur locally, but are of little importance. They might be termed canga deposits formed *in situ*, as they occur at the surface in itabirite areas where the leaching of silica has been sufficient to leave the residual material concentrated as iron ore. Frequently the iron oxide itself has been dissolved and redeposited in the form of limonite in pores and along

cracks as veins and stringers. In some places manganese oxide (psilomelane) is associated with this redeposited material.

The lenses and beds of carbonate rocks in the lower part of the Piracicaba formation in places contain a noticeable percentage of iron and manganese carbonates. On decomposition these form impure residual deposits of iron and manganese oxides.

Origin of the bedded ores.—From the structure and shape of the bedded ore deposits and their relation to the inclosing rocks it is necessary to conclude that they are original sedimentary deposits, laid down in a measure similar to limestone beds. They are simply a part of the iron-formation beds, similar in origin to the itabirite with which they are associated, except that they contain a smaller proportion of mechanical sediments, such as quartz sand. It is, therefore, best to explain the origin of the iron formation as a whole and note the differences in origin between the itabirite portions and the iron ore portions.

As has been stated, the iron formation is, in the main, an iron-oxide-bearing sandstone or quartzite. It varies in different portions from a rock consisting largely of quartz sand to one consisting of pure iron oxide. The siliceous portions of the iron formation are called itabirite, just as the siliceous portions of the Lake Superior iron formation are called ferruginous chert, jaspilite, or taconite. The portions rich in iron oxide are termed massive ore, powdery ore, or laminated ore, according to their texture. The line between the itabirite and the iron ore, especially the laminated ore, is arbitrary, depending on the percentage of metallic iron present. At the present state of the iron industry, iron formation with 50 per cent or more of metallic iron may be termed iron ore and iron formation containing below 50 per cent metallic iron itabirite. In the Brazilian iron formation there are no equivalents of the cherty iron carbonate and the greenalite rock of the Lake Superior district which in that region represent the original rock from which the iron ore, ferruginous chert, and associated rocks have been formed by weathering. In the Brazilian iron fields there has been no such alteration, the itabirite and iron ore being the original rocks.[1]

[1] C. K. Leith, and E. C. Harder, "The Hematite Ores of Brazil and a Comparison with the Hematite Ores of Lake Superior," *Econ. Geol.*, VI (1911) 670–86; E. C. Harder, "The 'Itabirite' Iron Ores of Brazil," *Econ. Geol.*, IX (1914), 101–11.

The iron formation has all the characteristics of a sedimentary deposit, although, just as in the rest of the Minas series, fossils have not been recognized in it. Portions are very massive and exhibit little or no evidence of bedding, as is frequently seen in massive quartzites, while other portions are thin-bedded or finely laminated, as shales and fine-grained sandstones. In the thinner-bedded portion the bedding or lamination planes are for the most

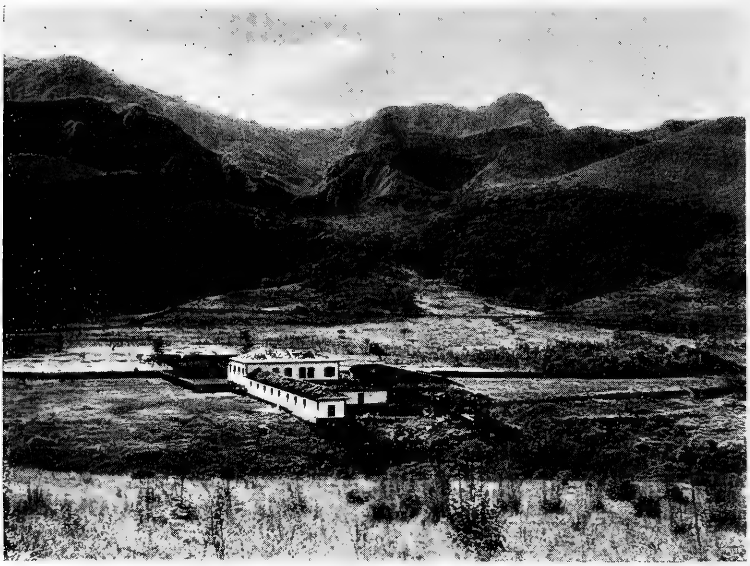

FIG. 16.—The Fazenda de Alegria. The highest peaks behind are of Caraça quartzite. The strong range of hills in front are of iron formation. To the left of the deep notch they are chiefly of itabirite; to the right of it dominantly of laminated iron ore.

part sharply defined. The stratification in places is rendered very conspicuous for the reason that in the itabirite portions of the formation the iron-oxide layers are so frequently separated by thin partings of white quartz sand. These sandy partings, as planes of weakness, control to a considerable extent the weathering and fracturing of the beds. While the main bulk of the iron formation consists of sandy iron-oxide-bearing beds, there frequently occur, interbedded with these, shaly or schistose iron-oxide-bearing beds

showing that, while the mechanical sediments consisted mainly of quartz sand, clay also was frequently deposited in certain localities.

Thus it is clear that in structure and general occurrence the iron formation is so like ordinary sediments that there is little doubt as to its sedimentary origin. It remains, however, to account for its peculiar and unique composition and also for its enormous mass. It is difficult to picture a source for such an enormous amount of iron oxide as is contained in the iron formation, and even if an adequate source of supply be found, it is not so easy to see how the iron oxide came to be deposited locally in masses hundreds of meters in thickness with less than $\frac{1}{2}$ of 1 per cent of siliceous sediments intermixed.

It is apparent from the composition of the Caraça quartzite that during the first stages of the sedimentary deposition the processes of decomposition greatly exceeded in effectiveness the processes of mechanical disruption, for, instead of arkose sediments, the Caraça quartzite is made up of the products of mature weathering. This may be ascribed to a long period of decomposition without transportation, resulting in the accumulation of a thick mantle of residual material on the land surface, or it may be ascribed to a warm, moist climate, or to extensive wave action with practically continuous sedimentation. The material of which the sediments are composed was derived from the decomposition of rocks of the basement complex, such as granite, gneiss, amphibolite, and crystalline schists, with some basic intrusives. The principal minerals associated in the make-up of these rocks are quartz, feldspar, biotite, hornblende, and some pyroxene, whose end products after weathering are quartz, kaolin, hydrated iron oxide, and some aluminum hydroxide. These four constituents, therefore, probably made up the bulk of the decomposition products.

During the sedimentation which followed the post-Archean erosion period the first material to be deposited in central Minas Geraes was quartz sand, from which it is natural to suppose that the finer products of the decay of the basement complex—the kaolin and hydrated iron oxide—were carried farther out to sea, while the coarser quartz sand was deposited nearer the shore. This process must have been long continued, for the Caraça

quartzite is thick. Following the Caraça quartzite is the Batatal schist, the finer material of which implies a slackening in the vigor of sedimentation, whether because the land from which the sediments were being derived was becoming lower, or because the source of the sediment was becoming more remote owing to an advancing shore line, or because of deepening of waters or shifting of currents. The Batatal schist is thin, indicating that this stage was not long enduring. It was followed by local flows of basic lava. And then came the deposition of the iron formation. The hypotheses to explain this remarkable formation naturally are of two sorts: (1) that the iron oxide was a mechanical sediment washed in like the mud of the schist and the sand of the quartzite; (2) that it was precipitated from solution through either chemical or biochemical agencies.

Following the hypothesis that the iron oxide was a mechanical sediment, the source for the iron oxide should naturally be the residual material resulting from the thorough weathering of the rocks of the basement complex. Besides this hydrated iron oxide, the residual material should contain kaolin and quartz in somewhat greater proportion than the iron oxide. Under any hypothesis the sediments which resulted in the Caraça quartzite and the Batatal schist came from the sorting of thoroughly weathered material by stream and wave action. The Caraça quartzite represents a time when only material of the coarser sort, the sand grains, could find rest in the area under consideration. The finer material was swept farther out to sea to find lodgment in deeper and quieter waters. With the lessening of the vigor of the agents of sedimentation came the thin Batatal schist. At this time fine mud was being washed into the sea and deposited, while the coarser sand was either not washed into the sea or was left still nearer the shore which may have changed its position in the meantime. With the kaolin much iron oxide must also have been washed into the sea and not a little of it deposited with the mud.

Thus far the process has been the ordinary one of decay of igneous rocks and the deposition of much of the resulting material as clastic sediment. The unusual feature to be explained is what very peculiar condition obtained to cause the deposition of ferric

hydroxide pretty generally over the region and in some places the building up of a series of beds of ferric hydroxide in a high state of purity which sometimes totals several hundred meters in thickness. The completeness of the separation, as evidenced by the fact that much of the ore now contains 69 per cent iron out of a possible 70 per cent for chemically pure ferric oxide, and the vast extent of the iron formation are very hard to understand. The iron oxide resulting from the decomposition of granite, gneiss, amphibolite, and crystalline schists should be in a more or less finely divided condition, and as such should not be very generally separated from the sand and clay by any sorting action of running water or ocean waves, except in so far as affected by a higher specific gravity which might result in small local accumulations of iron sands, or in the association of smaller particles of iron oxide with larger particles of other materials in the sediments. The iron oxide should be with the schists and quartzites. Great masses of iron oxide containing less than $\frac{1}{2}$ of 1 per cent of impurity do not seem possible as the result of mechanical separation, though it is not impossible that a small percentage of the iron oxide in the iron formation may have been derived thus.

The alternative hypothesis is that the iron oxide was precipitated from solution following the formation of limestones as an analogy. This escapes the serious difficulties of separation of materials. As in the case of the limestones, to account for the purity it is only necessary to suppose that the deposition of the iron compound occurred in clear seas where comparatively little clastic sediment was being brought in. While the thickness of iron formation is great, it is exceeded by many limestones.

It is the nature of the precipitation that leaves the widest field for speculation. There are two possibilities: (1) the iron may have been precipitated directly from solution by purely chemical means, or (2) it may have been abstracted from the water by organisms and deposited as a sediment. In either case the iron may be supposed to have been in solution chiefly in the form of ferrous carbonate and to have been thrown down as ferric hydroxide.

There are various chemical reactions which can cause the precipitation of ferric hydroxide from solutions of ferrous carbonate,

of which in nature oxidation of the carbonate with the resulting hydrolysis is perhaps the most likely possibility. But such a reaction should necessitate a notable quantity of iron compound in solution. Van Hise and Leith object that river and sea-waters do not contain the requisite amount of iron.[1] Their suggestion that the iron in the various iron formations of the Lake Superior region has come from associated basaltic lavas, either from the magmatic waters or from chemical reactions between the hot basic lavas and the sea-water,[2] hardly seems applicable to the Brazilian iron formation, since nothing in the nature of basic lavas has been found within the sedimentary series of the iron-ore district with the exception of the serpentinized remains of one small flow near Cattas Altas. This is very insignificant in extent in comparison with the iron formation, and furthermore extensive lenses of iron formation occur higher up within the Piracicaba schist with which there are no igneous rocks associated. In chemical precipitation there is likewise to be considered the fact that chemical reactions would be likely to produce other precipitates in addition to ferric hydroxide, such as aluminum hydroxide and calcium compounds, which would result in the formation of impure deposits of iron ore rather than in thick beds and lenses of pure ferric hydroxide.

The other possibility is that the ferric hydroxide was thrown down by organic action. It is now known that much of the bog iron ore being formed in lagoons at the present time is the result of the activity of a certain group of bacteria known as the iron bacteria. The iron bacteria include many individual species, of which the thread bacteria *Chlamydothrix*, *Gallionella*, *Spirophyllum*, *Crenothrix*, and *Clonothrix*, and the coccus form *Siderocapsa* have perhaps been most carefully studied.[3] While the different species have individual morphological peculiarities of their own, the type

[1] C. R. Van Hise and C. K. Leith, "The Geology of the Lake Superior Region," *Mon. 52*, *U.S. Geol. Surv.* (1911), pp. 503–6.

[2] *Op. cit.*, pp. 506–18.

[3] Hans Molisch, *Die Eisenbakterien*, p. 10, Jena, 1910; D. Ellis, "A Contribution to Our Knowledge of the Thread Bacteria," *Centralbl. für Bakt.*, Abt. II, Bd. 19 (1907), p. 502; Abt. II, Bd. 26 (1910), p. 321.

as a whole possesses certain general physiological characteristics. They all live in clear water, either standing or running water. Lieske states that he has never found any in turbid water, nor in waters containing a great deal of organic matter.[1] They live in waters containing iron compounds in solution which it is claimed by Winogradsky they utilize according to the following reaction: $2FeCO_3+O+3H_2O=2Fe(OH)_3+2CO_2$. Heat is liberated by this reaction, and this energy together with the carbon dioxide developed is utilized by the bacteria to sustain life.[2] Ferric hydroxide is left behind and may accumulate. Other investigators, like Molisch,[3] claim that ferrous compounds are not necessary for the physiological processes of these organisms and that organic compounds other than carbon dioxide must be present for their use. Nearly all agree, however, that their activity results in the accumulation of deposits of ferric hydroxide in many places. As the result of this activity the water pipes of cities where the water contains a considerable amount of ferrous carbonate have sometimes been completely closed.[4] That certain limonite deposits have been produced in this way is evidenced by the fact that in them large numbers of the sheaths of these bacteria have been found.[5] To quote from Lafar:

> The decomposing power of these organisms is very great, the amount of ferrous oxide oxidized by their cells being a high multiple of their own weight. This high chemical energy on the one hand, and the inexacting demands in the shape of food on the other, secure to these bacteria an important part in the economy of nature, the enormous deposits of ferruginous ocher and bog iron ore, and probably certain manganese ores as well, being the result of the activity of the iron bacteria.[6]

[1] Rudolf Lieske, "Beiträge zur Kenntnis der Physiologie von Spirophyllum ferrugineum (Ellis), einem typischen Eisenbakterium," *Jahrb. für wissenschaftliche Botanik*, XLIX (1911), 91–127.

[2] S. Winogradsky, "Über Eisenbakterien," *Botan. Zeitung*, Bd. 46 (1888), p. 261.

[3] Hans Molisch, *op. cit.*, p. 44.

[4] F. Lafar, *Technical Mycology*, I (1898), 361; also I (1910), 272.

[5] A. Fischer, *The Structure and Functions of Bacteria*, p. 69, tr. by A. Coppen Jones, Clarendon Press, Oxford, 1900.

[6] F. Lafar, cited by Van Hise and Leith, *op. cit.*, *Mon. 52*, *U.S. Geol. Surv.* (1911) p. 519.

Such a process as this would seem to offer a possible clue to the origin of the Itabira iron formation. As these bacteria thrive best in clear waters, the low proportion of clastic sediment in the formation is natural enough. If the deposition in general took place in lagoons and embayments and the thicker and more siliceous portions of the formation developed near the river mouths, in regions of currents, or other favorable localities, and the thinner, more uniform portions of the formation represent deposits formed

FIG. 17.—A view of the Peak of Conceição near Itabira de Matto Dentro, from the summit of Caué. The iron formation forms a continuous belt between Caué and Conceição.

farther off shore or in quiet waters near shore, the great and sudden variations in the thickness of the iron formation at various points may perhaps be accounted for. Much of the ferric hydroxide may have been formed as a flocculent precipitate in the sluggish river waters carrying little or no clastic sediment, and later deposited in a thick series of delta beds at the debouchures of the streams. Such may be the great accumulations of iron formation at Alegria, Gandarella, and elsewhere. Whatever the nature of this sedimenta-

tion, the field evidence appears to indicate that the deposition of the iron-oxide beds in these localities of unusual thickness took place with comparative rapidity.

While certain thin discontinuous layers of iron formation might be attributed to chemical percipitation, it is difficult to realize how great thicknesses over large areas could have been thus formed. Little is known concerning the concentration of iron compounds in solution necessary for chemical precipitation. It probably varies greatly with varying conditions, and the special conditions which would have to be assumed as causing the deposition of the Brazilian iron formation must have extended over large, as well as widely scattered, areas since this formation is not only found throughout a considerable area in Minas Geraes but also exists in the extreme western part of Brazil, nearly a thousand miles to the west. On the other hand, there is experimental evidence that bacteria do precipitate iron oxide out of very dilute solutions, and it is only necessary to assume the presence of large numbers of these micro-organisms in scattered localities to account for the presence of the iron formation. Unfortunately the metamorphism which the iron formation has suffered makes it impossible to recognize organic remains, if such were originally present. Recognizable bacterial remains, however, are stated to be extremely short-lived and it is difficult to identify them even in modern bog-iron-ore deposits. There is also to be considered the fact that nowhere in the ocean do we know of extensive deposits of ferric hydroxide being formed at the present time. This objection, however, is equally valid in case of ferric hydroxide precipitated either chemically or biochemically, and therefore must be considered under either hypothesis. In the formation of bog ores on continental areas bacteria are known to play an important part, but chemical precipitation is probably effective also, and it is difficult to say which plays the principal rôle. In spite of various uncertain factors, however, the deposition by micro-organisms seems more easily to explain the various unusual phases of the Brazilian iron-formation sedimentation, and we prefer to adopt this hypothesis, at the same time, however, realizing fully the inadequacy of our present knowledge concerning bacteria as a

geological agency, and the necessity for further investigation on this most important subject.

MANGANESE ORES

With the exception of gold, manganese ore is the most important of the metalliferous mineral products of Brazil at the present time. Two mines in central Minas Geraes, the Morro da Mina mine north of Lafayette, and the Wigg mine east of Miguel Burnier, are in continuous operation, while several smaller mines produce ore intermittently. Among these are the Rodeio mine near Kilometer 508, east of Miguel Burnier on the Ouro Preto branch of the Central Railroad of Brazil, the Cocuruto mine southwest of Lafayette, and the Queluz das Minas mine near the Morro da Mina mine north of Lafayette. Many abandoned and inactive mines occur in the general vicinity of the mines mentioned above.

The manganese deposits of central Minas Geraes may be separated into two distinct classes: (1) those occurring in the basement complex and (2) those occurring in the overlying sediments. The deposits found in the region around Lafayette belong to the first class, while those occurring along the Ouro Preto branch railway east of Miguel Burnier are of the second class. The centers of these two districts, having distinct types of deposits, are not more than 25 kilometers apart, and some of the mines in the one can be seen from the other. The deposits have been described in detail by Derby[1] and by Scott,[2] so that only a general outline need be given here in order to show their relation to the general stratigraphy.

The manganese deposits which are found in the basement complex consist of large irregular masses of manganese oxide inclosed in, or bounded by, gneiss, granite, or crystalline schist. Individual masses such as that at Morro da Mina may be more than a hundred meters in their longer diameter. While irregular in shape, they are usually somewhat elongated, suggesting lenses. They occur

[1] O. A. Derby, "On the Manganese Ore Deposits of the Queluz (Lafayette) District, Minas Geraes," *Am. Jour. Sci.*, XII (1901), 18–32; "On the Original Type of Manganese Ore Deposits of the Queluz District, Brazil," *ibid.*, XXV (1908), 213–16.

[2] H. K. Scott, "The Manganese Ores of Brazil," *Jour. Iron and Steel Inst.* (1900), p. 179.

scattered through the basement complex without any apparent regularity, but most of them appear to have either gneiss or crystalline schist on one or both bounding walls.

The manganese oxide composing these lenses is usually in the amorphous form, occurring mainly as psilomelane and wad, though pyrolusite also is found with these. According to detailed studies made by Dr. Derby,[1] it appears that these oxides are surface decomposition products of other manganese minerals which have in one or two cases been encountered below the zone of oxidation. Of these minerals the principal ones are the manganese silicates, tephroite and spessartite, and with these occur rhodochrosite, the manganese carbonate, and sparingly rhodonite, another manganese silicate. These minerals occur intimately intermixed in varying proportions, one being more abundant in one place and another elsewhere, and together they form a reddish manganese silicate and carbonate rock. The relation of the manganese rock to the inclosing crystalline rocks has not been definitely determined; it may be interlayered with the gneiss or crystalline schist or perhaps it is intrusive into them.

From the one or two instances noted it is judged that all the manganese oxide deposits in the areas of basement complex are surface oxidation products of such masses of manganese silicate and carbonate rock. In many of the deposits where the original rock has not been encountered, the oxide ores have textures which are duplicated in the manganese silicate and carbonate rock elsewhere, and therefore suggest a similar origin. During the process of decomposition more or less solution and redeposition takes place, with the result that certain portions of a deposit are composed of high-grade manganese oxide, while other portions contain admixtures of other products of decomposition, such as clay and quartz sand. Most of the ore is hard, but soft material, mainly wad and pyrolusite, also occurs abundantly, being irregularly intermixed with it.

Manganese ores associated with igneous rocks, such as those described above, occur abundantly in India and are also found locally in the eastern United States.

[1] O. A. Derby, *Am. Jour. Sci.*, XXV (1908), 215–16.

The manganese deposits in the sedimentary series occur as definite beds associated with the iron formation. The principal bed, that on which the Wigg mine is situated, is 3 or 4 kilometers in length, and at its maximum reaches a thickness of over 2 meters. It strikes in an east-west direction parallel to the strike of the inclosing sediments and corresponds with them in dip, making it apparent that the manganese ore bed was laid down as a sedimentary bed just as the inclosing rock. The bed at the Wigg mine is bounded on one side by soft itabirite, with a contact zone of mixed soft hematite and manganese oxide, and on the other side by a ferruginous schist associated with the iron formation.

The manganese bed at the Rodeio mine is of smaller horizontal extent but of greater thickness than that at the Wigg mine and shows less definitely its relation to the inclosing rocks. In the case of both of these deposits, beds of carbonate rocks consisting of a mixture of calcium, magnesium, iron, and manganese carbonates are found in the vicinity. These, however, occur at different horizons from the manganese beds.

The manganese ores associated with the sedimentary rocks consist of finely crystalline or amorphous manganese oxides, probably largely a mixture of pyrolusite and psilomelane. From their occurrence it must be assumed that they are similar in origin to the associated rocks, that is, that they are original sedimentary deposits of manganese oxide which have been somewhat altered and recrystallized by subsequent metamorphism. The source of the manganese is doubtful, but it may very well have been derived from deposits of manganese ore in the basement complex such as now occur to the south near Lafayette. Decomposition of such deposits may have yielded a large amount of residual manganese oxide which was worked over, transported, and deposited as beds or lenses in the sedimentary series. Their origin would thus be very similar to that of the iron ores of central Minas Geraes with which they are closely associated.

Sedimentary manganese ore beds similar to those of central Minas Geraes are found in many places on other continents, probably the best known of them being those in northern Arkansas, those in central Chile, those in the Caucasus near the Russian boundary, and those in western Arabia.

GOLD

Gold was known to occur in Brazil even in the early days of its settlement and it was the search for this metal which brought many of the explorers and settlers to the new country. However, not until the beginning of the eighteenth century was the great gold field of Brazil, that occurring in central Minas Geraes, actively worked. It was at this time that the towns of Sabara, Marianna, and Ouro Preto, the latter then known as Villa Rica, were founded

FIG. 18.—Ouro Preto, former capital of Minas Geraes. A city built by the gold-mining industry. The hills on all sides show the results of placer mining.

in rapid succession. A period of great excitement and activity followed, which continued in waves as districts were extended and new ones discovered, almost to the middle of the nineteenth century. From that time on the industry rapidly decreased in importance until 1889 when gold mining as a general industry practically ceased as a result of the abolition of slavery. At the present time there are only two important gold mines operated in Minas Geraes. These are the Morro Velho mine at Villa Nova de Lima near Bello Horizonte and the Passagem mine at Passagem near Ouro Preto.

Nearly all the early workings were shallow open cuts or shafts, or short tunnels. Later development led to the operation of a few deep mines, such as the Descoberto mine near Sabara, the Gongo Socco and São Bento mines between Caethé and Santa Barbara, the Pary mine near São Francisco, and the Santa Anna and Maquiné mines near Marianna. The shallow workings cover large areas and are widely distributed in central Minas, groups of them being found in the vicinity of nearly all the small towns and villages, for in most cases it was the gold mines that caused the founding of the villages. Some of the areas of old abandoned workings cover many hundreds of acres.

Gold occurs in Minas Geraes in three different associations: (1) in quartz or sulphide veins, (2) disseminated in the iron formation and in the canga derived from it, and (3) in stream gravels.

Gold-bearing quartz veins occur with different relations; some of them are strike veins, others occur along bedding or schistosity planes; some are long and continuous, others short and lens-like. In some places groups of short parallel or intersecting veins are found, while elsewhere single isolated veins of considerable extent occur.

Quartz or sulphide veins containing gold may occur in the basement complex or in any of the sedimentary formations. The Morro Velho mine at Villa Nova de Lima is operating a large vein in the Piracicaba schist on which they have descended for a vertical distance of more than 1,700 meters, deeper than any other gold mine in the world. The Passagem mine near Ouro Preto is working on an irregular bedding vein impregnating a thin layer of Batatal schist between the Caraça quartzite and the Itabira iron formation. In the old workings near Ouro Preto quartz veins are found cutting the Caraça quartzite and the Itabira iron formation. Near Cattas Altas old workings are found in the basement complex and here also quartz veins occur. Many other examples might be given of quartz veins found in the various formations.

The mineral veins of the district for the most part come under two general heads: (1) ordinary quartz veins, and (2) veins of magmatic origin. To the first class belong certain quartz-hematite veins occurring in the iron formation, and the numerous barren

quartz veins found throughout the district, while to the second class belong the quartz-feldspar pegmatites and the gold-bearing quartz, or quartz-sulphide veins.

The quartz-hematite veins in the iron formation consist of quartz in which flakes of specular hematite occur. The hematite may be found in large, irregular, curved flakes along cracks, or it may be imbedded within the solid quartz. These and various barren quartz veins are of little or no importance as gold-bearing veins. They are probably ordinary circulating water depositions, as the quartz does not contain inclusions of monazite, zircon, rutile, garnet, or xenotime, some of which are nearly always present in the quartz in pegmatite veins and in igneous rocks.[1] It is possible, however, that this evidence may not be conclusive and that by more detailed study some of these veins may be found to be hot-water depositions.

The quartz-feldspar pegmatite veins have already been mentioned in the description of the rocks of the basement complex. They consist of quartz, feldspar, and muscovite with which other minerals, such as beryl, columbite, tourmaline, etc., are sometimes associated in minor quantities. They occur only in rocks of the basement complex and are not gold-bearing.

The gold-bearing quartz and sulphide veins are the important sources of gold in the district. They occur in various formations and differ somewhat in mineral composition in different localities. In some places free gold occurs in quartz, sulphides being inconspicuous, while elsewhere the gold occurs in arsenopyrite, pyrrhotite, or pyrite associated with a variable amount of quartz. Other minerals occurring in places in or near these veins are calcite, cyanite, biotite, garnet, oligoclase, tourmaline, albite, siderite, muscovite, and others.

At the Passagem mine[2] the principal sulphide minerals are arsenopyrite, pyrrhotite, and pyrite, occurring in a gangue of quartz, or quartz and decomposed oligoclase, strongly impregnated with tourmaline. The oligoclase is altered to calcite and white mica. With these occur in varying abundance the other minerals

[1] Specimens examined by Dr. Derby.

[2] A. J. Bensusan, "The Passagem Mine and Works," *Inst. Min. and Met.*, Twentieth Session, October 19, 1910.

mentioned above. Dr. Derby[1] has made a study of the genesis of the mineral and ore deposition at this mine and has established three successive stages of mineralization. He believes that the original quartz-oligoclase deposition was of a pegmatitic nature, the material being derived from some intrusive igneous mass which, however, has not yet been discovered. For although nowhere in the district has any evidence been found of igneous intrusions into the sedimentary series, the nature of the mineralization is such as to leave no doubt as to its magmatic origin. During the deposition of the quartz and oligoclase it appears that garnet (andradite), biotite, cyanite and staurolite were developed locally along the contact, these minerals occurring in the country rock along the border of the vein. Crystals of apatite are associated sparingly with the garnet. Graphite also is associated with the contact minerals, being found along shearing planes.

The second stage of mineralization, according to Dr. Derby, consisted in the introduction of tourmaline along cracks and as impregnations into the earlier minerals.[2] This stage was very pronounced, tourmaline being one of the principal gangue minerals. Its introduction was closely followed by the third stage, namely the sulphide mineralization, by which were introduced arsenopyrite, pyrrhotite, and pyrite. During the introduction of both the tourmaline and sulphides, the alteration of the oligoclase to calcite and white mica took place. Both tourmaline and sulphides occur with the white mica and calcite in the decomposed oligoclase masses, their abundance depending on the degree of alteration. There is evidence that the tourmaline and sulphide solutions attacked the contact minerals as well as the vein material. The biotite is decomposed in places, and tourmaline and sulphides are commonly associated with the contact minerals.

In the vicinity of the gold-bearing lode, but not directly connected with it, are certain geodes with the following mineral association: calcite, siderite, albite, quartz and muscovite. The

[1] O. A. Derby, "On the Mineralization of the Gold Bearing Lode of Passagem, Minas Geraes, Brazil," *Am. Jour. Sci.*, XXXII (1911), 185–90.

[2] *Op. cit.*, p. 190.

origin of these and their relation to the minerals of the gold-bearing lode has not yet been determined.

Summarizing the foregoing statements we may separate the various minerals into the following groups:

Original Vein Minerals	Contact Minerals	Later Vein Minerals		Geode Minerals
Quartz	Garnet	Tourmaline	Arsenopyrite	Calcite (surface etched)
Oligoclase (largely altered to white mica and calcite)	Biotite (locally altered)		Pyrrhotite	Siderite (surface dissolved)
	Staurolite		Pyrite	Albite
	Cyanite	Secondary Minerals		Quartz
	Apatite	Sericite		Muscovite (with sericite)
	Graphite (associated with contact minerals)	Calcite		

The famous Morro Velho mine at Villa Nova de Lima, which has been the greatest gold-producer in Minas Geraes, is operating on a vein in the Piracicaba schist. This vein is parallel to the bedding which here dips toward the southeast at an angle of about 45°. It has an average width of 5 to 10 meters, and this notable width continues with some variation to the greatest depth yet reached, which in 1912 was more than 1,600 meters below the surface. The lode reaches a length of more than 150 meters along the strike.

While the country rock is the Piracicaba schist, the gangue material is a fine-textured mixture of carbonates (siderite, dolomite, and calcite) with quartz, and, according to Dr. Derby, a small amount of albite.[1] The ore consists principally of pyrrhotite with smaller amounts of arsenopyrite, pyrite, and chalcopyrite. A notable feature of the vein is its constancy in width, mineralization, and values through the numerous levels down to the lowest one yet opened.[2]

[1] O. A. Derby, "Notes on Brazilian Gold Ores," *Trans. Am. Inst. Mining Eng.*, XXXIII (1902), p. 282–87.

[2] For a brief history and general statistics of these mines see: H. K. Scott, "The Gold-field of the State of Minas Geraes, Brazil," *Trans. Am. Inst. Min. Eng.*, XXXIII (1902), 406–44.

The gold disseminated through the iron formation and resulting canga formed the basis for most of the gold-mining operations in the early years. A considerable proportion of the iron formation being soft, it was washed almost as easily as stream gravels and, therefore, was attractive to the early miners.

Gold occurs as the native element disseminated in the Itabira iron formation, as well as in the iron-formation lenses of the Piracicaba schist. It may be found in hard, or soft ore, or in itabirite, but generally it occurs in the soft schistose phase of the iron formation known as "jacutinga" (see p. 389).[1] Most of the old workings are in soft itabirite, or in soft, powdery ore, doubtless because of the facility of operating in these formations. The gold is very unequally disseminated within the iron-formation belts, there being large areas of iron formation in which little or no gold occurs, and other places where it is found gathered in rich pockets. There is no regularity in the distribution of the gold-bearing portions of the iron formation and no apparent physical condition, unless it be difference in porosity, which would cause the localization in the places where it occurs. Because of the irregularity of distribution and the disseminated nature of the gold, this type of deposits has never formed the basis of large individual operations. The deposits were worked during the time of slavery because of the cheapness of labor, but when slavery was abolished the workings were abandoned.

The gold occurring in the iron formation is of varying coarseness, from fine dust, hardly visible to the naked eye, to fragments several centimeters in length. Even large masses are reported to have been found. The larger pieces have a porous, spongy texture and contain intermixed particles of iron oxide and quartz. The particles of gold are generally elongated or platy in form, owing to deposition along the lamination planes of the iron formation.

Most of the gold occurring in the iron formation is very pure and contains but very little alloyed silver or other metals. Locally, however, there are occurrences of what is known as *ouro branco* (white gold), which consists of an alloy of gold and palladium.

[1] E. Hussak, "O Palladio e a Platina no Brazil," *Annaes da Escola de Minas de Ouro Preto*, N. 8 (1906), p. 96, tr. by Miguel A. R. Lisboa, and Manoel A. R. Lisboa.

This alloy has been encountered in a number of places, but is nowhere very abundant.

Quartz in veins or lenses is very common in the iron formation, and in places contains gold. It is very often found where disseminated gold occurs, and it is possible that some genetic relation exists between the gold in the quartz and that disseminated in the iron formation. As the masses of vein quartz in this association are generally small, irregular, and discontinuous, it seems possible that, because of the porous nature of the formation, the gold-bearing waters did not follow regular channels but impregnated the iron formation for considerable distances on both sides of the main lines of flow and thus deposited their gold in a more or less disseminated condition. At any rate, from its form and occurrence it is certain that the gold is a secondary concentration within the iron formation.

Gold occurs in the gravels of most streams rising in, or flowing through, the gold-bearing district (Fig. 19). These occurrences are simply placer deposits whose gold has been derived in part from gold-bearing quartz veins, but mainly from the great masses of iron formation, the disseminated gold of which, because of its widespread occurrence, offers abundant opportunity for concentration along streams.

At present small gold-washing operations are conducted by natives along many of the streams at the end of each rainy season, since the flood waters caused by the heavy tropical rains bring new material down the valleys each year.

DIAMONDS

The principal diamond fields of Brazil are those of central Minas Geraes near Diamantina, of Western Minas Geraes near Bagagem, and of Bahia near the towns of Sincora, Lencoẽs, and Jacobina. Of these the Diamantina field is the oldest and best known and has yielded most of the best Brazilian diamonds. The Bagagem field has yielded most of the large diamonds found in Brazil, while the Bahia field supplies nearly all the world's demand for carbonados.

The Diamantina field is located in the Serra do Espinhaço quartzite belt, the principal workings lying north, east, and south

of the town of Diamantina. They occur on portions of the main tableland which forms the watershed between the Rio Jequitinhonha and the Rio São Francisco, as well as on highlands between the upper tributaries of the Rio Jequitinhonha.

The Serro do Espinhaço in this portion of its extent is a belt of uneven, rolling tableland from which rise numerous small, irregular quartzite knobs and ridges. The plateau is deeply cut into by the

FIG. 10.—The Rio Gualaxo at Antonio Pereira. The scene of much placer mining in the past and where small gold-washing operations are still conducted by natives at the end of each rainy season. The prominent peak is that of Frazão, a lens of hard quartzite in the Piracicaba schist

headwaters of the Jequitinhonha which form great, rocky gorges. The gorges are, however, comparatively recent, and by far the larger portion of the region still remains as broad stretches of uplands known in Brazil as *chapadas*. Above the general level of the uplands in the eastern part of the district rises an old monadnock, the isolated peak of Itambé, which is one of the highest points in Minas Geraes. The uplands have a general elevation varying from

1,200 meters to 1,300 meters above sea-level. The Jequitinhonha where it leaves the district has an elevation of about 900 meters, while Itambé rises above the uplands to an elevation of 2,000 meters above the sea; thus the greatest relief in the district is about 1,100 meters.

The tablelands or *chapadas* are of great interest, both geologically and commercially, for it was on them that the diamonds were

FIG. 20.—The diamond-washing of Serrinho near Curralinho. In the distance is the monadnock of Itambé.

first concentrated and it is on them that the principal deposits are now found. The *chapadas* are, for the most part, underlain by Caraça quartzite, dipping uniformly but generally at low angles to the east. On many of them, especially those near the actual watershed of the Serra do Espinhaço, there are remnants of what were once more extensive deposits of gravel, sand, and clay. These have been called the Diamantina conglomerate. The remnants lie on the eroded surface of the Caraça quartzite, and when examined carefully it is found that most of them occur in irregular trough-like depressions, the bottom and walls of which consist

of quartzite (Fig. 21). Many of these troughs reach a depth of 30 meters or more, though some of the deeper ones scarcely exceed 100 meters in width (Fig. 22). No determination has been made of their maximum width, while as to their original longitudinal extent nothing is known. From their irregular occurrence and their relation to the underlying quartzite it is presumed that these deposits are for the most part the result of stream deposition.

Fig. 21.—The pit of Serrinho. A shallow depression in the Caraça quartzite which was partially filled with diamond-bearing conglomerate. The conglomerate and beautifully white kaolin are seen in the bottom of the excavation. The final washing operations to recover the diamonds are conducted in the *batêa,* or jig, under the thatched roof.

This view is further strengthened by the nature of the materials composing them and the relation of the different kinds of material to each other. By detailed mapping it might be possible to connect all these remnants so as to reproduce the old drainage network as it was when the conglomerate was deposited.

The gravel, sand, and clay occur associated in the deposits as beds, lenses, or masses, without any regularity. Locally beds of pure

clay, or clay slightly intermixed with sand, occur, while elsewhere abundant pebbles are scattered through clay or sand matrices. The nature of the pebbles in different parts of the district varies, but the characteristic and dominant pebbles consist of quartzite, undoubtedly derived from the Caraça formation. Associated with these are pebbles of iron formation, quartz, schist, diorite, amphibolite, and other rocks, the association varying in different

FIG. 22.—The trench at Cadette's mine, northwest of São João da Chapada

localities. A characteristic but less abundant pebble in these deposits is the diamond.

Where exposed at the surface the conglomerate is very hard and indurated, but beneath the surface both matrix and pebbles, especially those of quartzite, are in many places soft and friable. This softening of the quartzite pebbles bears evidence of considerable age of the conglomerate deposits.

In many respects these deposits are strikingly similar to certain deposits of gravel, sand, and clay which occur along the Appalachian region in the United States. In Georgia these contain bauxite;

farther north in scattered localities they contain concentration deposits of brown iron ore and manganese ore. In the United States these deposits have been referred to the Tertiary or late Cretaceous.

From the great predominance of quartzite pebbles it must be supposed that the materials composing these soft conglomerate deposits came largely from the Caraça quartzite. The immediate source of the igneous pebbles present in some localities, however, is less certain. These may once have been laid down in conglomerate beds in the Caraça quartzite and have been worked over again later, or they may have been derived from the rocks of the igneous area adjacent to the Espinhaço sedimentary belt. The Caraça formation in this district is a well-consolidated quartzite of medium-grained uniform texture. Conglomerate beds are rare, and though scattered pebbles are frequent, these are almost invariably of vein quartz. There is a possibility, however, that pebbles of igneous rock may be present locally.

Accepting the supposition that the material composing the conglomerate was derived largely from the Caraça quartzite, one would naturally look for the origin of the diamonds in the same formation. In the Diamantina district, diamonds have never to our knowledge been found in the Caraça quartzite. It is, however, possible that they may occur so widely scattered that it is only by repeated concentration such as has occurred in the conglomerate deposits that they become noticeable. In the Grão Mogul district of northern Minas Geraes about 200 kilometers to the northeast of Diamantina, diamonds are reported to be found in hard quartzite[1] from which they are freed by blasting. However, it is not known whether this is the Caraça quartzite or whether it belongs to some other formation.

In conformity with the generally accepted theory of their origin, the diamonds may have been derived originally from intrusions of igneous rock at a distance from the present areas of concentration. Disintegration and decomposition of the inclosing rocks would have resulted in the freeing of the diamonds, leaving them

[1] O. A. Derby, "Modes of Occurrence of the Diamond in Brazil," *Am. Jour. Sci.*, 3d Ser., XXIV, 39.

in residual accumulations subject to removal by streams and other subaerial agencies. If in part transported to the sea during the deposition of the Caraça formation, they should be found probably in association with quartz sand and such pebbles of vein quartz as resisted the abrasian of stream action. Being the hardest of all minerals and the most resistant to decomposition, it is not surprising that the diamonds still retain their crystalline faces when most other minerals originally present, except quartz, have disappeared.

When in pre-Devonian times the sedimentary series was elevated, consolidated, and metamorphosed, and the processes of erosion had commenced, the diamonds began gradually one by one to be freed from the quartzite and collected in the stream deposits along with other minerals from the same source. This continued for long ages, and while other minerals were disintegrating and decomposing, the diamonds remained intact and became more and more concentrated. Even after a general peneplain level was reached, this process probably continued and was only interrupted by renewed elevation resulting in the present gorges.

The foregoing discussion is based on the hypothesis that the diamonds at one stage in their history were incorporated in the Caraça quartzite. The alternative hypothesis is that the majority of them never have been deposited in the quartzite, but that they were brought in by streams directly from igneous areas some distance from the present diamond fields. An objection to this hypothesis is that so few igneous pebbles occur in the conglomerate, and that over large areas pebbles of igneous rocks are entirely absent. However, it is possible that, on account of their predisposition to rapid decay, they might have disappeared in the slow, shifting process on the peneplain surface. But one wonders why, if such diamond-bearing igneous-rock areas exist, not concealed by the sedimentary rocks, they have not been discovered up to the present time. If the diamonds were brought in from a long distance the *chapada* diamond-bearing gravels should necessarily occur along a few main drainage courses and not be irregularly scattered over a wide area, while if they came from igneous areas close by, the diamonds should have been discovered in streams flowing from these areas. But in the present state of knowledge it

seems best to leave the question whether the diamonds in these *chapada* conglomerate deposits have, for the most part, been derived directly from igneous rocks of the Archean complex, or whether, coming in any case from that source originally, there occurred an intermediate stage of incorporation in the Caraça quartzite—an open one.

Perhaps the most noted of the diamond mines of this district is that of São João da Chapada, which has been rather fully described by Dr. Derby.[1] It is located some 30 kilometers northwest of Diamantina on the tableland which represents the old base-leveled surface. Sunk in the solid quartzite of the tableland is a steep-walled trench, perhaps 500 meters long and now about 30 meters in depth (Fig. 23). This trench has the appearance of being the channel of an old stream which became completely filled with residual clay, sand, and gravel when the surrounding region was close to the base level. This alluvial-filled trench has been cleaned out again as the deposits of clay, sand, and gravel have been washed away in the process of diamond mining.

Dr. Derby in several of his papers has been inclined to the hypothesis that the diamonds of São João da Chapada are vein minerals,[2] derived from a pegmatite vein of which there seems to be some evidence in the cut. Certain masses of pure-white kaolin containing nests of large and beautiful crystals of quartz, which have never been exposed to the wear of running water, look much like the decay products of a pegmatite vein. Unfortunately at the time of our visit in 1912 the mine had not been worked for twenty years, so that definite and reliable information was not easily obtained, but from the accounts given by some of the people at São João da Chapada and the published papers which touch upon this point, it would seem that the diamonds were far more abundant and characteristic of the waterworn conglomerates, especially where pebbles of higher specific gravity (iron ore, itabirite, etc.) were

[1] O. A. Derby, "Brazilian Evidence on the Genesis of the Diamond," *Jour. Geol.* VI (1898), 121–46; "On the Association of Argillaceous Rocks with Quartz Veins in the Region of Diamantina, Brazil," *Am. Jour. Sci.*, 4th Ser., VII 343–56.

[2] O. A. Derby, "Modes of Occurrence of the Diamond in Brazil," *Am. Jour. Sci.* 3d Ser., XXIV, 34–42; "The Genesis of the Diamond," *Science*, IX (1887), 57–58.

common, than of those particular clayey portions which suggest a pegmatitic source. Because of this the São João da Chapada occurrence does not seem essentially different from the other diamond-bearing conglomerate deposits. A pegmatitic dike, being less resistant than the surrounding quartzite, may have determined the location of the drainage channel in the first place, and because of its ready yielding gave rise to the steep-sided trench which would

FIG. 23.—The famous diamond mine of São João da Chapada. The conglomerate-filled trench in the quartzite has now been largely re-excavated, the undisturbed quartzite appearing at various points in both walls and in the bottom of the excavation. This mine has produced some of the finest of the Brazilian diamonds.

otherwise seem peculiar in a peneplained surface. But we believe that if such pegmatitic material was present it was only a contributing condition, and that the diamonds came with the material that was washed into the trench filling it up.

Since their deposition with the residual base-level gravels which now constitute the *chapada* or upland deposits, several reconcentrations of the diamonds have occurred, as the region has suffered various changes of attitude. Upon the rejuvenation of the drainage

the streams began dissecting the old nearly base-leveled plain, first cutting through the surface decomposition products and associated fluviatile deposits, and then attacking the underlying quartzite. As the *chapada* deposits were being removed in this erosive process, the softer and lighter materials were being worn finer and washed away, while the heavier diamonds remained behind and became further concentrated in the stream gravels of the immediate

Fig. 24.—Diamond mining at Cadette's mine. The clay and materials of lesser specific gravity are washed away by running water while the diamonds and other heavy pebbles remain behind and become concentrated.

vicinity. At several stages during this erosion there were periods when the downward cutting of the streams was checked and they began to widen their valley bottoms and deposit material over them. Then active cutting began again, leaving gravel-covered terraces on the valley slopes. Much material from these several gravel terraces as well as from the diamondiferous conglomerate of high-level *chapadas* has since been washed down the present steep

valley slope into the Jequitinhonha River. This has resulted in the following distribution of the diamond-bearing deposits. Numerous remnants of *chapada* deposits still occur upon the plateau in areas as yet undissected by the rejuvenated streams. Gravel deposits of later age occur lower down on valley terraces at various elevations above the present stream beds. Gravel and sand deposits of still more recent age occur along the present stream

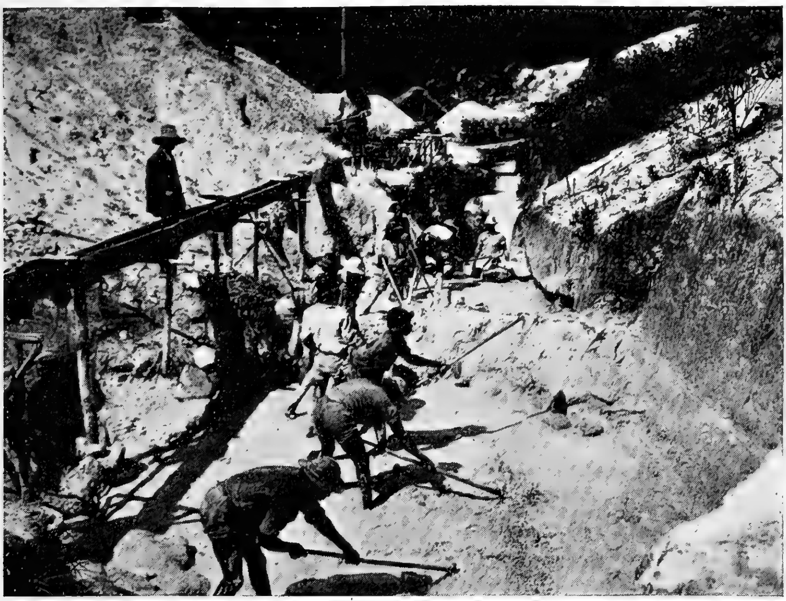

Fig. 25.—Stirring up the mortar-like mass of diamond-bearing clay and gravel

bottoms. All these deposits contain diamonds in greater or less abundance.

The present-day practice for the recovery of the diamonds is simply a furtherance of the process of concentration which has been going on in nature. When a considerable quantity of soft kaolin and loose gravel from the diamondiferous conglomeratic deposits has been accumulated in the bottom of the trench, or mine, a stream of water is conducted through the trench (Fig. 24). As the running water passes over this loose material, natives armed with a sort of

hoe keep the mass constantly stirred up, so that it has a general consistency and appearance not unlike thin mortar (Fig. 25). The running water carries away the clay and materials of lesser specific gravity, while the coarse portions of the gravel, and the minerals of higher specific gravity such as the diamond, fragments of hematite, etc., remain behind. Later, after screening out the coarser pebbles, the diamonds are picked out of the final concentrate on the *batêa*, or washing sieve, by hand.

The diamonds of the Diamantina district generally show dodecahedral crystallization with rounded faces. Many of those in the *chapada* deposits have rough faces and a dark grayish-green coating, but the majority have smooth or striated faces. Some of the stones in the present river gravels show surface corrasion.

The majority of the diamonds mined in the district vary in size from $\frac{1}{4}$ carat to 4 or 5 carats, though occasionally stones are found weighing up to 10 carats. The largest diamonds ever found in the Diamantina district, according to Dr. Derby,[1] weighed less than 100 carats, and very few have been found weighing over 50 carats. The largest of the Brazilian diamonds—the "Star of the South," the "Star of Minas," and the "Dresden" diamonds—have all come from the Bagagem district in southwestern Minas Geraes.

In brilliancy the Diamantina diamonds exceed the South African diamonds, and there is a smaller percentage of "off-colored" stones found. Beautiful bluish-white stones are abundant, and stones of other colors such as lemon-yellow, cognac-brown, rose, green, and blue are occasionally found.

We are indebted to Professor T. C. Chamberlin, Professor C. K. Leith, and Professor Eliot Blackwelder for criticisms and suggestions, and to Dr. O. A. Derby for much information.

[1] O. A. Derby, "A Notable Brazilian Diamond," *Am. Jour. Sci.*, XXXII (1911), 191.

THE STRENGTH OF THE EARTH'S CRUST

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PART VIII. PHYSICAL CONDITIONS CONTROLLING THE NATURE OF LITHOSPHERE AND ASTHENOSPHERE

SECTION A[1]

RELATIONS BETWEEN RIGIDITY, STRENGTH, AND IGNEOUS ACTIVITY

INTRODUCTION AND SUMMARY

The experiments of F. D. Adams have demonstrated that under combined pressures and temperatures equal to those existing at a depth of eleven miles granite is about seven times stronger than at the surface, showing that, up to at least certain limits, the strength of the crust increases downward. The measurements of tidal deformation of the earth and of the variations of latitude concur furthermore, in proving that the rigidity of the earth as a whole is greater than that of steel. The transmission of earthquake vibrations of a transverse nature through the earth shows, not only that the earth is solid and rigid throughout, but that, on the whole, rigidity increases with depth. In none of these lines of investigation is there any clear suggestion of the existence of a thick shell of weakness—an asthenosphere.

On the other hand, the conclusion that broad areas of the crust rest in approximate isostatic equilibrium seems to imply the

[1] Section B of Part VIII, on "Relations with Other Fields of Geophysics," will be published in the succeeding number of this *Journal*.

existence of a subcrustal zone with but little strength, readily yielding under vertical loads when these are of such breadth that the strains resulting from them are not confined and absorbed within the strong outer crust.

There has thus been developed a paradox, an apparent conflict of evidence which becomes more insistent of explanation with continued accumulation of proofs of high rigidity from the domain of geophysics and of proofs of regional isostasy from the equally precise field of geodesy.

In the consideration of such broad problems, the hope of an ultimate definitive solution rests upon the use of the method of multiple working hypotheses. The surety, significance, and breadth of application of the facts must be established. By these the various hypotheses must be tested and molded. All hypotheses must be kept for further consideration provided they are not positively excluded. In the complexity of relationships there is commonly a complexity of cause, and hypotheses which seem at first to be mutually exclusive may be found to co-operate in giving a completer explanation. Those which originally appeared antagonistic may thus come to be seen as participating and dividing the field of cause between them. A paradox often points to this kind of a conclusion.

To pass to the particular problem of the relations of isostasy to the physical conditions of the earth's interior; the hypothesis developed in this study—of the existence of a zone of weakness underlying a zone of strength—must not be regarded at present as the only available hypothesis. Searching investigation must be carried forward to see if other and possibly antagonistic hypotheses cannot be developed which will equally well co-ordinate and explain the facts. Even if true in the main, it is likely, as has been the case with other hypotheses, that time will show that in certain directions it has been carried too far. Such testing, however, can best be done by others, and after the implications of this hypothesis are seen. In this part, the concluding article of this series, a discussion had best be given of the lines of adjustment by which the hypothesis here favored may be brought into harmony with other fields of geophysical evidence. With this understanding of the relation of

the present investigation to the method of multiple working hypotheses, examination will be made of the paradox which has been drawn between certain conceptions from other lines of investigation and those drawn from this study of crustal strength.

Having given this introductory presentation on what is conceived to be a judicial point of view, we may turn to a review of the conclusions reached in this article. It is pointed out that rigidity is strictly a measure of stiffness; whereas a very different quality, the limit of elastic yielding, or the beginning of flow, is the measure of strength. But mass flowage may take place in a number of quite different ways, according to the nature of the solid and the environing physical and chemical conditions. The elastic limit and hence the strength will differ in the same solid according to the mode of yielding. Four modes may be here enumerated in what is thought to be their order of increasing importance, the fourth mode being that which is conceived as operative especially in the asthenosphere, and serving to maintain the condition of approximate regional isostasy.

First, flowage may take place rapidly by true plastic or molecular flow, as with lead or white-hot iron, the solid, when stressed well beyond the elastic limit, behaving like a viscous fluid. It is not thought that the terrestrial deformations are often carried on with a rapidity which requires true plastic yielding. In fact, under such rapid stresses as those produced by earthquakes and tides it is not improbable that the strength of the earth may progressively increase with depth.

As a second and quite different mode, deformation may take place by molar as distinct from molecular shear. In the zone of fracture this is manifested in jointing and faulting and is emphasized as distinct from rock flowage, but where the fracturing becomes so closely spaced as to result in slicing of individual minerals it passes under the category of granulation. Where carried on at depth there is always some degree of cementation by recrystallization. Deformation by such close-grained fracture without complete loss of cohesion is classed as rock flowage. It is thought to be developed to some degree within the lithosphere, especially by great horizontally compressive forces, but is not

regarded here as the mode of deep rock flowage involved in the isostatic readjustment of unfolded tracts.

Thirdly, flowage may take place in some minerals, as calcite and ice, by gliding upon the cleavage planes. But such gliding is not regarded as the mode by which the foliated rocks are developed. It requires furthermore far greater force than that which is given by the departures from isostatic equilibrium.

A fourth mode of rock flowage is by recrystallization. It is the chief factor, as Van Hise has shown, in the deformation of the crystalline foliates. It is thought that this is also the method by which the asthenosphere yields and that a readiness of recrystallization under unbalanced stresses of a permanent nature is the cause of the weakness of the asthenosphere.

The vibratory forces transmitted as earthquake shocks and those due to tidal strain, from this standpoint, are both rapid. Under such conditions the asthenosphere could show high order of strength. It is argued that the ease of recrystallization under constant strain becomes more marked the nearer the temperature approaches to that of fusion, or to express it better from the physico-chemical standpoint, the nearer the temperature approaches to the mutual solution point of the constitutents involved. The result is that at such temperatures the rigidity may be high and not greatly different from that at low temperatures, but for permanent stresses the elastic limit becomes low. The movement of continental glaciers with a low surface gradient, accomplished by recrystallization, illustrates the condition which it would appear exists to even a higher degree within the asthenosphere.

This conclusion carries with it the idea that within the lithosphere the temperature is in general considerably below that of fusion; whereas below, in the thick zone of weakness, the temperature must lie close to, or at, that of molten rocks. For fusion there is needed, however, the energy necessary to supply the latent heat and volume expansion. Unless this energy is supplied, the asthenosphere remains solid rock, but the least accession of internal heat, or relief from external pressure, will generate a proportionate amount of magma, diffused as liquid throughout the solid. To gather into reservoirs temporarily molten, the magma must con-

verge by rising, analogous to the draining of melted water from glaciers; uniting, as rivulets unite into rivers, and rivers discharge into lakes. No continuous lava stratum or large reservoirs of lava could, under the terms of this hypothesis, be expected to exist within the asthenosphere. Its very weakness would prevent it from acting as a containing vessel for holding large volumes of any fluid which, for any cause such as a lower specific gravity of the fluid phase, would tend to rise. The evidence of earthquake vibrations and of resistance to tidal deformation further supports the view that the asthenosphere is not a liquid or even a truly viscous zone. On the other hand, only in the lithosphere would be found the strength needed for the storage of magma in volumes until the limit of its strength as a containing vessel was reached.

Partly guided by observation upon the metamorphic rocks, partly by theories of the nature of deformation at great depths, the argument leads to conclusions on the mode of yielding within the different levels of the crust. First, the outermost zone is observed to be a zone of fracture, weak in comparison with the thick zone below. This, the second zone, is the zone of strength and yields by flowage, but flowage which is characterized by granulation as the dominant, by recrystallization as the subordinate, mode. The expenditure of energy for a given deformation is here a maximum. In the third zone, the asthenosphere, on the contrary, flowage is conceived as taking place with but little expenditure of energy, by a ready recrystallization at the temperature of primary crystallization of magmas. Those contorted granite-gneisses seen especially in the Archean rocks, which are regarded as deformed during the final stages of crystallization, exhibit locally in the outer crust the conditions which it appears may permanently prevail within the asthenosphere.

SECTION A

RELATIONS BETWEEN RIGIDITY, STRENGTH AND IGNEOUS ACTIVITY

DISTINCTIONS IN PHYSICAL PROPERTIES RELATED TO STRENGTH

Elasticity is of two natures: that of volume and that of form. The first is possessed by matter in either the gaseous, liquid, or solid state; the second is possessed by solids only and is associated

with rigidity and strength. Up to a certain degree of strain known as the elastic limit, elasticity of form in the ideal solid is perfect and is expressed by the law that the change of form, or strain, is directly proportional to the load applied, or stress. This load may be maintained indefinitely and, except for a slight relaxation, the solid shows no further yielding. Upon the removal of the load there is an elastic return to the original form, but the very last stages of the recovery are slow. The elastic nature of the whole earth, in regard to both volume and form, is shown by its capacity to transmit the several kinds of earthquake vibrations. The permanence and perfection of the elasticity of form is also implied in the power of the crust to carry loads up to certain limits for times reaching into geologic periods without exhibiting progressive viscous yielding.

Beyond the elastic limit, elasticity ceases to be perfect, and a permanent change of form occurs. This relieves part of the stress and reduces the strain to within the elastic limit. The change of form may be by rupture, in which event the strength of the body is destroyed. It may be by plastic flow, in which case the strength may be increased or decreased. Wrought iron, for example, becomes somewhat stronger as a result of forging. Granite on being mashed into gneiss becomes somewhat weaker because of the development of weaker minerals, especially the micas. In the crust of the earth, except for the outer few miles, flowage takes place without probably much change in the mineral composition and consequently in the strength of the rock. Deformation will continue as long as the stress is maintained well above the elastic limit, but upon the cessation of movement there may still remain residual stresses up to the elastic limit. If the residual stresses over broad areas are small, it may be because the development of weaker structures, such as folds or zones of igneous injection, has eased the strains.

Failure by flow brings in the distinction between viscosity and plasticity. These are often used, even by physical geologists, as merely synonymous terms, but there is a real distinction which should be noted. Fluids are viscous to a small or large degree and can have no elasticity of form. Viscous flow must, however, overcome internal friction and requires time for its accomplishment.

With long time even a minute force will cause even a very viscous fluid to flow. Solids, on the other hand, possess elasticity of form, and below the elastic limit can hold shearing stresses indefinitely. Above it they may flow and in so doing exhibit plasticity. The phenomenon differs from viscosity in that the force must rise to a certain magnitude before any gliding between molecules begins. The crust, then, is plastic but not viscous.

Although the theoretical distinction between plasticity and viscosity is clear, recognition must be given to conditions where the two states merge. This is especially true for undercooled glasses. A glass in its molecular organization is a liquid and yet it possesses definite elastic moduli and elastic limit. From this standpoint of elasticity the glass, therefore, is a solid. Upon rise of temperature there is, however, no absorption of latent heat to mark a change of state, the elastic limit gradually lowers, disappears for prolonged stresses, and elastically, the substance passes by gradation from a solid to a liquid. The existence of these transition cases should not, however, be permitted to obscure the real distinctions between solid and liquid.

The crust yields as a plastic solid to forces which strain it beyond its elastic limit. But the solid flowage which this implies may be either by distortion of crystals or by recrystallization. The first is familiar for rapidly applied forces, requires comparatively great stress, and corresponds to the usual conception of plasticity. The crystalline rocks make us familiar, however, with the idea of mass plasticity by recrystallization. This is plasticity, but in a somewhat different sense from that which is usually conveyed by the term.

The degree of elasticity which a substance may exhibit is a different property from the elastic limit. A bar of wrought iron one square inch in cross-section will be elongated one part in 28,000,000 by a tensile stress of one pound. A similar bar of glass would be elongated one part in 10,500,000, more than twice as much. These ratios measure the degree of elasticity under tensile or compressive stresses and differ for each substance. The figure is known as Young's modulus of elasticity. A substance may be highly elastic, that is, have a high modulus of elasticity, as cast iron, or glass,

and yet be brittle because of a low elastic limit under rapid tensile stress, combined with lack of plasticity at ordinary temperatures. At temperatures sufficiently high, the modulus is not greatly different, but the elastic limit is still lower. The substance is now, however, plastic, rather than brittle, since plasticity is greatly increased; but a rapid strain, exceeding the rapidity with which plastic deformation can take place, may still produce fracture. Another substance, such as rubber, may have a low modulus of elasticity and yet a relatively high elastic limit.

Among similar substances under similar physical conditions there is, however, a definite association of these properties which for the metals is brought out well in a tabulation by Johnston and L. H. Adams.[1] It is shown for a class of substances, such as the metals, that the modulus of elasticity, the hardness, the tensile strength, and the elastic limit all, so far as the data are given, occur in the same order; so that of two metals that which has the higher elastic limit is the higher also in the other qualities. From this association there results a ready mental confusion between rigidity and strength. The one, however, denotes the degree of resistance to distortion from a unit-shearing stress and gives the modulus of ridigity. The other is measured by the elastic limit. As an example of the confusion between these two different properties, it is known that the earth as a whole is more rigid than steel. This to many would appear to mean that it was stronger than steel. Earthquake waves show that the earth becomes progressively more incompressible and more rigid with depth. This might be held as evidence against the existence of a thick sphere of weakness, the asthenosphere. High incompressibility and high rigidity are not, however, direct testimony of strength, and it is the purpose of the next topic to show under what conditions a solid may be very rigid and yet very weak.

CONDITIONS FAVORING ASSOCIATION OF HIGH RIGIDITY WITH LOW ELASTIC LIMIT

Alpine glaciers as well as the Alpine-like margins of the Greenland ice sheet move much more rapidly in the summer than in the

[1] "On the Effect of High Pressures on the Physical and Chemical Behavior of Solids," *Amer. Jour. Sci.*, XXXV (1913), 220.

winter, a phenomenon to be accounted for by the rate of recrystallization. The parts of an ice crystal which are subjected to shear and compression have the melting-point lowered. They melt, discharge the strain, and refreeze. In the winter the general temperature is reduced, and a greater strain is necessary to bring the melting-point down to the lower temperature. Until local melting is produced the ice behaves like any other crystalline solid, as a substance possessing elasticity of form. Beyond that point it exhibits plasticity and behaves in some respects like a very viscous fluid. In other respects, however, it exhibits properties quite distinct from that of the usual conception of mere plastic flow, since in the testing machine, or on the walls of a crevasse, ice will resist strong shearing strains, and yet the glacier as a whole yields and flows slowly under a moderate pressure-difference as shown by the low gradient of its upper surface. Glacial motion appears to take place, therefore, by the solution and growth of crystals, not by a true viscous flow. The solid and crystalline nature throughout as opposed to viscous fluidity is furthermore shown, as Chamberlin has noted, in the power of the glacial ice to shove over and abrade its floor and to ride up slopes. Chamberlin adds that a dry glacier is a rigid glacier. A dry glacier is necessarily cold, and a cold glacier is necessarily dry.[1]

With ice subjected to slowly applied forces the elastic limit is consequently dependent upon the point of yielding by recrystallization. We thus see an intimate relationship between temperature and variation in the elastic limit, the elastic limit for ice being greater for low temperatures than for high temperatures. But the modulus of rigidity, on the contrary, measures the elastic change of form for unit-shearing force, change of form not accompanied by crystallization, but marked by a capacity to spring back to the original form upon the removal of the stress.

T. W. Richards, in his studies on the compressibility of solids, notes that they are almost as compressible and voluminous at absolute zero as at ordinary temperatures. Under this conception

[1] "A Contribution to the Theory of Glacial Motion," *Decennial Publications of the University of Chicago*, IX, 203, 204 (1904); Chamberlin and Salisbury, *Geology*, I (1904), 305.

the molecules, taken as equivalent to their spheres of influence, are in actual contact and suffer mutual compression owing to the attraction of cohesion. The influence of heat is relatively unimportant in determining the density of a solid. The atoms are in most cases even more compressed and distorted by the converging force of chemical affinity than are the molecules by cohesion. This corresponds with the fact that substances of small atomic volume are on the whole more incompressible than those of greater atomic volume. For the more incompressible substances also the decrease in compressibility with added load is relatively little, suggesting that they are already greatly compressed by the forces of chemical affinity.[1] The effect of heat serves only to distend slightly the spheres of influence of the atoms, so long as the substance is in the solid state. Rise of temperature to near the melting-point, as long as there is not a softening by the development of incipient liquidity, should, according to these views, change the elastic properties but slightly. The greatly lessened strength of ice near the melting-point, as expressed in the freedom of regelation, is not, following these ideas, connected with the slightly lessened incompressibility and rigidity. For many substances the problem is complicated, however, by changes of molecular state with changes in temperature and pressure. This is especially true of ice when subjected to extreme ranges in temperature and pressure, as has been shown by Bridgeman; for ordinary glacial ice, however, we deal with but a single state.

For ice at a temperature of $-7.^{\circ}03$ C. the compressibility has been determined for pressures ranging approximately between 100 and 500 atmospheres. It is found to possess, according to Richards and Speyers, about one-fourth of the compressibility of water at neighboring temperatures and about five times the compressibility of glass.[2] But glass possesses a compressibility between that of acidic and basic holocrystalline igneous rocks. Ice may be taken then as about three or four times as compressible as granite.

[1] "The Present Aspect of the Hypothesis of Compressible Atoms," *Am. Chem. Soc. Jour.*, XXXVI (1914), 2417–39.

[2] T. W. Richards and C. L. Speyers, "The Compressibility of Ice," *Am. Chem. Soc. Jour.*, XXXVI (1914), 491–94.

Now the modulus of rigidity is related to the modulus of compressibility by means of a formula involving Poisson's ratio.[1] This ratio varies for each substance, but for rocks, for iron and steel, and probably for ice, it lies between 0.2 and 0.3 in value, so that in general the rigidity of these substances can be judged roughly by their degree of incompressibility. Consequently it is seen that glacier ice at temperatures such as those which prevail in the body of the moving glacier possesses a degree of incompressibility and rigidity which, if these elastic constants were measures of its strength, would make it wholly incapable of motion on such gradients as are observed. This can be made more obvious by some quantitative statements. Granite and similar rocks, for example, can stand permanently in steep cliffs to heights of thousands of feet. They constitute mountain ranges whose height and steepness are limited entirely by the forces of erosion on the one hand and the strength of the asthenosphere on the other. The cliffs could be very much higher and the mountains much more lofty before glacier-like flow at the base of the mountain mass would occur. In fact, with a compressive strength of 25,000 pounds per square inch, a rectangular block of granite could stand as a vertical wall 22,000 feet high, and of indefinite breadth, without yielding of the base. With a sloping face and supported by spurs such as occur in nature, the height of the granite mass could become considerably greater. For parallel mountain ranges of harmonic form and gentle slopes resting upon a foundation whose compressive strength to indefinite depths was 25,000 pounds per square inch, the mountain crests could stand eleven miles above the valley bottoms before the maximum stress-difference would reach this limit. Even then, if the slopes were as low as those of a continental ice sheet, the failure would not take place by flowage of the mountains laterally into the valleys, but by a vertical settling of the mountains and a vertical upwarping of the valleys. The lateral, plastic flow would be at some depth in the earth. If the asthenosphere were indefinitely rigid, granite mountains of sufficiently gentle slope

[1] Let P = Poisson's ratio; C, the modulus of rigidity; D, the modulus of compressibility. Then $C=\frac{3}{2}\,\frac{1-2P}{1+P}\,D$.

could rise to indefinite heights. This is because the depth of maximum stress-difference would lie at about one-sixth of the wave-length below the mean level of the surface. With increasing wave-length the height of the waves could accordingly be greater without increasing the stress-difference at the trough-line of the waves. The gradient would, however, have to become more gentle; in other words, the amplitude would have to increase at a lesser rate than the wave-length. If the strength of ice were measured by its rigidity it could stand permanently in masses one-third or one-fourth as steep and high as these theoretic limits for granite mountains, without failure by plastic flow. Yet, on the contrary, the great ice fields spread out by flowage of their bases, although their surfaces possess very gentle gradients. The distinction between strength and rigidity in the movement of glaciers is thus clear. The strength of glaciers is limited by the amount of the stress-differences needed to produce slow movement by recrystallization.

Johnston and L. H. Adams have applied this theory of yielding, well known as an explanation of glacial motion, to all plastic flow, and argue that even for those substances, such as the metals and rocks, in which cubic compressibility raises the melting-point, shear greatly lowers it for the parts under stress.[1] They argue from a physico-chemical basis that the most plausible explanation for flow in metals is that the shearing strain is great enough on individual points to produce a change of phase of individual molecules from solid to liquid, even at ordinary temperatures.

Apart from theory as to its explanation, the phenomenon of welding of iron shows for high temperatures a low elastic limit and ready passage beyond into plastic flow. For iron and steel, furthermore, the influence of temperature upon the rigidity has been investigated. Pisati gives the following equations in which n is the value of the modulus of rigidity for temperature t.[2] For iron—

$$n_t = 811 \times 10^6 (1 - .000{,}206t - .000{,}000{,}19t^2 + .000{,}000{,}001{,}1t^3),$$

[1] "On the Effect of High Pressures on the Physical and Chemical Behavior of Solids," *Am. Jour. Sci.*, XXXV (1914), 205–53.

[2] *Smithsonian Physical Tables* (1904), p. 76.

for steel—

$$n_t = 829 \times 10^6(1 - .000,187t - .000,000,59t^2 + .000,000,000,9t^3).$$

This equation for iron gives a minimum modulus at 314° C. equal to 95 per cent of the modulus at 0° C. For steel the minimum value occurs at 342° C. and is 90 per cent of the modulus at zero. At 528 °C. iron has the same modulus as at 0° C. and at 890° C. steel has the same modulus as at zero. Doubtless 1000° C. is above the limits of the data from which these formulas were derived. For this temperature they may consequently give inaccurate results, but it is of interest to note that the curve gives a modulus of rigidity for iron at that temperature 1.7 times that at 0° C. and for steel 1.1 times that at 0° C. The extrapolation prevents attaching quantitative value to these figures, but the qualitative conclusion may be reached that iron and steel at high temperatures do not exhibit less rigidity than they possess at lower temperatures. It is obvious, however, that above a certain temperature the elastic limit becomes very low, as shown by the capacity for forging, and for strains beyond this limit deformation takes place by plastic flow. That it is not merely incipient fusion is suggested by the maintenance of a crystalline condition through the process of deformation. The subject for iron is doubtless complicated by the fact that iron passes through more than one solid molecular state in being heated up to fusion. Presumably then the equation given for the relation of rigidity to temperature can only be a first approximation to the actual changes.

Let the attention be given next to the crystalline rocks which were once deep-seated and, owing to subjacent batholithic invasion, attained their crystallization at exalted temperatures. It is observed that, although the rock masses have been extensively deformed, the individual crystals have regrown during the process so as to possess compact boundaries, and an internal constitution nearly free from strain. The interpretation is that the deformations due to geologic forces were so slow and the rocks were so saturated with crystallizing agents at high temperatures that recrystallization could nearly keep pace with the deformation, even for temperatures below the range of plasticity. As understood by

the students of anamorphism, the process has depended upon a readier solution of the molecules under shearing stress than of those free from such stress—solution carried on by means of the relatively minute proportion of gaseous crystallizers which were present through the rock mass. Such crystallizers doubtless facilitate the process. They form, in fact, solutions with the rock which may be regarded as mixtures with very low fusion points. But, theoretically, as the temperatures approach those of general fusion the need of such crystallizers diminishes. Moderate shearing stresses can thus liquefy the parts upon which they act and a process analogous to glacial motion sets in for solid rock. The strength of highly heated rock appears then dependent upon the amount by which the temperature is below the melting-point. That zone of the earth which is very weak may then be regarded as approximately at the temperature of fusion. To transform the solid into liquid there is needed only the energy required for latent heat and increase of volume. The proportion of liquid which is generated will vary directly with the amount of heat supplied or the amount of hydrostatic pressure removed. Magma, consequently, can be generated in this zone more readily than above in the zone of strength; but it will not be in reservoirs; rather will it be in its place of origin disseminated through the rock mass like water standing in a porous sandstone.[1]

ANALOGIES BETWEEN ASTHENOSPHERIC ROCK AND GLACIAL ICE

The theory of the asthenosphere as here presented is seen to have important relations to other branches of geology. The zone of weakness becomes especially the generator of magmas; the

[1] Recently the writer has learned from Mr. Bailey Willis of an unpublished paper which he gave some years ago to the Geological Society of Washington, in which he outlined his views of the nature of crustal thrusts as illustrated by the Appalachians. In that, and more recently, as a result of studies in the Alps and Andes, he has come to entertain the view that the zone of compensation, the lithosphere, shears over the zone below through the agency of molecular or mass fusion. Deep-seated horizontal shear and igneous intrusion he thus holds have important associations with orogenic movements. We have thus arrived independently at somewhat the same view of the nature of the zone of weakness. The part which recrystallization may play in promoting such movement is suggested in his *Research in China*, Vol. II (1907), "Systematic Geology," pp. 130, 131.

pyrosphere has its roots in the asthenosphere. But in the attempt to frame a logical picture of the processes which determine the ascent of magmas, the question arises how diffused liquid matter is drained away, rising and uniting at higher levels into magma reservoirs, temporarily molten, and the direct source of the igneous activity exhibited within the outermost crust. A deductive picture is as follows—one whose truth cannot be tested directly, but only by its general agreement with our understanding.

At the place of origin, liquid of an andesitic or basaltic nature will come to honeycomb the rock. The content of gases is presumably sufficiently high to reduce the viscosity. The liquid will then become able to transmit hydrostatic pressures, and, although comprising only a part of the rock mass, will constitute a continuous column of considerable height. Then becomes possible the second stage, the draining upward and the convergence of the fluid rock. Gravity is the ultimate cause, as in the downward movement of waters, but here the fluid, being lighter than its surroundings, tends to move upward. An explanation of this draining process has been given by Lane.[1] In a gas-saturated rock an excess of gas, or liquid and gas, would have the power of opening fissures at any depth in the zone of flow without the necessity for the existence of any tensile stress in the walls, or competence in the walls to maintain an open cavity. All that is necessary is that the excess pressure in the rising wedge of gas should be stronger than the cohesion of the rock. The fissure becomes filled with gas and fluid of lesser density than the solid rock of the walls. Consequently, the pressure transmitted from below is greater than the resisting pressure in the walls. This insinuating power, owing to the hydrostatic head due to the lesser gravity of the wedge, becomes greater the higher the wedge rises above the source, until near the surface the action may become violent and rapid. Daly also has outlined a theory of mechanism for the injection of abyssal wedges of magma into the upper crust.[2]

[1] "Geologic Activity of the Earth's Originally Absorbed Gases," *Geological Society of America Bull.*, V (1894), 259–80.

[2] *Am. Jour. Sci.*, XXII (1906), 195–216; *Igneous Rocks and Their Origin* (1914), chap. ix.

His theory is constructed however for action within the roof of a magmatic substratum. Daly postulates a zone of tension, but the mechanism suggested by Lane does not require this and would seem to apply better to the region of generation of magmas, for there the cubic compression is enormous, it is not a zone which has been subjected to cooling, and therefore it is difficult to conceive of the cause of a system of tensile stresses within the asthenosphere.

We are now prepared to draw a closer analogy between the physical conditions of the asthenosphere and those of a glacier, noting the likenesses and also the unlikenesses. In the summer, in the case of Alpine glaciers, heat is supplied to the surface of the glacier until it is warmed to the melting-point, and part of the ice absorbs the amount required by the change of state and passes into water. This trickles along the surface until a fissure is met and the water sinks by force of gravity toward the bottom. Near the snout of the glacier the temperature of this deeper part may thus be more dependent upon this convection than upon direct conduction through the ice. The winter freezing tends to chill the deeper ice and slow its motion, but during the summer the descending water tends to raise the temperature toward the freezing-point. In parts of the glacier where the heating is more effective than the cooling, the waters drill channels and gather at the base of the glacier into streams, reaching finally the outer world. The descent, and gathering, and englacial flow of glacial waters is analogous to the rise and convergence of streamlets of molten rock.

In order to account for lateral mass movement within the asthenosphere, an imperfection of isostasy between continental interior and ocean basin, giving an isostatic gradient or slope toward the continental interior, seems a necessary postulate. Such a gradient is so low that it has not yet been sifted from those irregularities of mass which are owing to the strength of the outer crust, or, it may be, obscured by great compressive bowings of the crust. This isostatic gradient, the slope required to generate movement within the asthenosphere, is far lower than that of the surface of a continental ice sheet. The failures of the analogy between the asthenospheric and glacial states are then as instructive as the agreements. The glacier is thin and broad. Friction on the bot-

tom is excessive and the motion requires more internal work. Much of its mass is permanently well below the freezing temperature. These are the factors which determine the steepness of the surface gradient. The asthenosphere by contrast should be deep and the differential motions within it necessary to satisfy isostasy would be correspondingly small. The temperature through a wide zone should be that of fusion under the hydrostatic pressures prevailing. The solid rock should be sodden with occluded gases, giving mobility to the growing fluid and ready to play their part in assisting recrystallization. Such a physical condition, as long as there is a continuous solid, would exhibit perfect elasticity and high rigidity during the passage of transverse vibrations, yet would slowly yield to prolonged shearing stresses, even though these were very small in amount.

RELATIONS OF IGNEOUS ACTIVITY TO ASTHENOSPHERE AND LITHOSPHERE

The argument has led to the view that the asthenosphere is a region where the temperature curve becomes tangent to the fusion curve, but that a condition of solidity is maintained by the recurrent elimination of that material which becomes molten. The importance of such a process, maintaining the solidity of the earth, has been dwelt upon by Chamberlin, especially as accounting for the overwhelming igneous activity of Archeozoic time. In lessened measure it applies to all later times as well.

Becker has held that the bottom of the zone of isostatic compensation is the depth at which the temperature curve approached nearest to the fusion curve, and he was the first to connect in this way the geodetic evidence with a temperature relation.[1] But Becker does not conceive of actual, permanent contact of the two curves as occurring, and took the depth of nearest approach to fusion as 122 km. This follows from Hayford's hypothesis of a uniform distribution for isostatic compensation, but in the present work there has been found reason for believing that compensation fades out through a greater depth; the strength, as measured by

[1] "Age of a Cooling Globe in Which the Initial Temperature Increases Directly as the Distance from the Surface," *Science*, XXVII (1908), 227–33, 392; "The Age of the Earth," *Smithsonian Miscellaneous Collections*, LVI, No. 6 (1910), 1–28.

the existence of stress-difference, through a depth greater still. The beginning of permanent contact of the two curves, if this is the cause of the disappearance of strength, should be as much as 300 km. deep and extend through some hundreds of kilometers.

A rectilinear projection downward of the temperature gradient observed at the surface would reach the fusion temperature of rocks at a depth of about 50 km. There must be consequently a marked curvature of the temperature gradient if the temperature and fusion curves do not meet short of 300 km. This curvature implies that near the surface there is either a greater quantity of heat flowing outward by conduction or that the conductivity of rock near the surface is very greatly decreased. But such a very great decrease in conductivity making for a higher temperature gradient finds no supporting evidence. On the other hand, a greater outward flow by conduction of heat near the surface may be due to the continued generation of heat by radioactivity to a greater degree than below; or also to a rise of magmas from the asthenosphere. Magmas which never reach the surface would bring heat by a convective process directly into the outer crust. From there the heat, slowly diffused upward by conduction, would increase the temperature gradient in the outermost part of the lithosphere. It is this factor especially which the argument of the present chapter emphasizes.

It is only within the present generation that general recognition has been given to the intrusive nature of the abyssal igneous rocks. They are now generally regarded as risen from the depths. Their action has been to break through and engulf the foundations of the ancient crust.

This process of batholithic invasion seems to be recurrent and widespread, though rise into the outer crust is restricted to the crises of diastrophism and usually reaches levels exposed to erosion only along the lines of mountain systems. The stores of heat brought up from the greater depths would be held in the crust, especially in its deeper parts, for geologic ages, blurring out in the course of time by conduction and creating a false appearance of heat lingering from an initial molten state, a resemblance increased by the added veil of new heat of radio-active origin mantling the ancient stores.

The temperature gradient under this view should naturally vary widely from place to place and from time to time. Igneous activity is the effective means by which heat is brought up from depths which on account of the slowness of conduction would be otherwise thermally isolated from the outer crust. Offset against this, cooling by conduction advances downward from the surface, dissipating not only the heat of local radio-active origin but that excess rising from the depths. The heat of the crust is not then a continually ebbing residuum from a primal molten state, but represents rather an oscillating ebb and flow, one of the balances of nature maintained through geologic time.

If this view be true—that the invasive igneous rocks have been an important factor in determining the amount and distribution of heat in the crust—it is doubtful if any sound arguments can be derived from the study of the present gradients as to the initial temperature. This conclusion is similar to the change of viewpoint in other lines of geology. It was once thought that the composition of the present atmosphere and the character of present climates were steps in a simple and continuous series of changes passing from primal conditions to a future in which the water would be absorbed into the earth and its surface transformed into a frozen desert. Now, however, it is generally recognized that since the earliest known times the surface conditions have been in a state of oscillating equilibrium. The argument of this section leads toward the view that this is true for the physical conditions within lithosphere and asthenosphere also.

[*To be concluded*]

THE BURIED ROCK SURFACE AND PRE-GLACIAL RIVER VALLEYS OF MINNEAPOLIS AND VICINITY

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University of Minnesota

CONTENTS

I. INTRODUCTION AND SUMMARY

The rock formations underlying the mantle of glacial drift which covers the region around Minneapolis have been studied and described in detail by N. H. Winchell,[1] Warren Upham,[2] C. W. Hall,[3] F. W. Sardeson,[4] C. P. Berkey,[5] and others. The character of the surface of this buried rock floor has been known only in a general way as revealed by local post-glacial erosion and, in part, by the present surface topography. The definite relations of this buried topography to the present one; the undulations of its surface; and the depth and courses of the buried river channels have been known only at scattered points where data have become available through occasional well borings or other artificial or natural excavations. In the investigations herein described an attempt has been made: (1) to show in detail, so far as available data will justify, the exact nature of the surface of the bed rock; (2) to determine the depth

[1] *Final Report Geol. and Nat. Hist. Survey of Minn.*

[2] *Ibid.*

[3] "Geology and Underground Waters of Southern Minnesota": Water Supply *Paper, U.S. Geol. Survey No. 256.*

[4] "Galena Series," *Bull. Geol. Society of America*, XVIII (1907), 179–94.

[5] "Paleography of Saint Peter Time," *ibid.*, XVII (1906), 229–50.

of the drift at all points within the city of Minneapolis; and (3) to trace the courses of the glacial and pre-glacial river channels, now filled with glacial débris, throughout the city. The results of these investigations are shown on the topographic map on which surface and solid rock is contoured and in the series of geologic structure sections which accompany this paper and to which frequent reference will be made in the following discussion.

The results show the following essential facts. In pre-glacial time the region around Minneapolis was dissected by a large river (referred to tentatively as the pre-glacial Mississippi) and its tributaries, which cut deep valleys into the rock. At a later time, during the Glacial period, the surface of the rock included in the areas between these old stream valleys was also deeply eroded by ice gouging and planation. While the effect of this glacial erosion was to produce a generally flat rock surface above the valleys, there were numerous irregularities of surface developed, many of which have never before been accurately delineated. The ancient valleys became filled with glacial débris which was brought down by the ice and left behind as a thick mantle covering the entire region when the ice sheet receded to the north. Not only were the old valleys choked up with glacial drift, but the entire rock surface was buried. The depth of this drift has been determined at as many points in the city as possible and has been found to vary from 0 to 250 feet. It is thickest in the old buried valleys and thinnest where post-glacial erosion has been active. The depth, width, and courses of the old buried valleys have been determined and are shown in contour on the accompanying map which gives a picture of the buried rock surface throughout the city. In constructing the map (Fig. 1), the surface topography has been superimposed upon this buried topography so that the depth to bedrock at any point may be readily determined from the map by taking the difference between the elevations shown by the surface contours (solid lines) and the bedrock contours (broken lines). The large depression which is the site of Powderhorn Lake was found to be due to the head of a small buried valley in this locality. Five secondary valleys which were tributary to the main buried river valley have been traced. These are shown on the map. Finally, it has been shown that the position of these buried

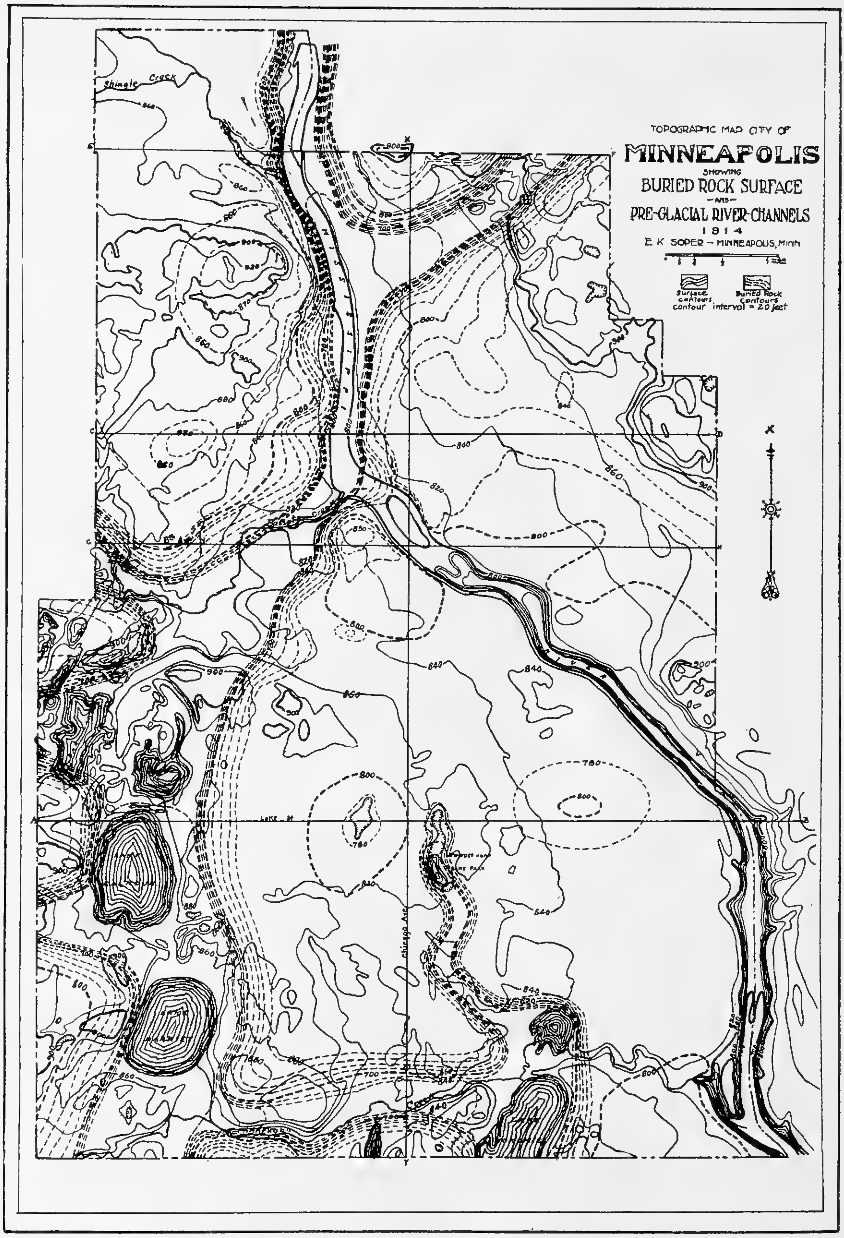
TOPOGRAPHIC MAP CITY OF
MINNEAPOLIS
SHOWING
BURIED ROCK SURFACE
—AND—
PRE-GLACIAL RIVER-CHANNELS
1914
E. K. SOPER – MINNEAPOLIS, MINN
Surface contours
Buried Rock Contours
contour interval = 20 feet
Shingle Creek
Lake St
Chicago Ave

FIG. 1

valleys should be a controlling factor in the choice of types of foundations for large buildings or other structures erected within the zone.

The data used in constructing the map and sections have been obtained from many sources, but chiefly from records of excavations for sewers and water mains and from hundreds of well records supplied by the well drillers operating in the vicinity of Minneapolis. The surface topography is partly taken from the topographic maps of the United States Geological Survey covering the region around Minneapolis and St. Paul, and partly compiled from elevations supplied by the city engineer, and from field work done in the spring of 1914.

The writer desires to express his thanks to all those who have given assistance in gathering data for this paper. Special acknowledgment is made to City Engineer F. W. Cappelen, Dr. F. W. Sardeson, and to the engineers in the Sewer and Water departments of the City Engineer's office; to Mr. A. C. Godward, engineer for the Minneapolis Board of Park Commissioners; Mr. James G. Houghton, city building inspector; Mr. J. F. McCarthy, Mr. E. F. McCarthy, the S. Swenson Artesian Well Co., and Mr. Max Renner.

II. ROCK FORMATIONS BENEATH THE DRIFT

The rock formations existing in the vicinity of Minneapolis, with the relative position, thickness, and characteristics of each, are shown in Table I.

Only four of the formations given in the table may constitute the rock surface immediately beneath the glacial drift in Minneapolis. These are (1) the Decorah shale; (2) Platteville limestone; (3) St. Peter sandstone; and (4) Oneota dolomite. The existence of the latter in direct contact with the drift has been determined at only two or three localities within the city limits, where deep wells have penetrated the drift. The localities have been found to lie over the deepest part of the main buried river channel, as shown by the deep well records. Therefore it seems probable that the pre-glacial Mississippi had cut entirely through the overlying formations and flowed upon a rock bed of Oneota dolomite from a point near the mouth of Bassett Creek in North Minneapolis southward to the city limits and beyond.

All four of the formations, each of which constitutes the first rock beneath the drift at different places, originally extended over the entire area of the city and far beyond. Before pre-glacial valleys were carved the rock now topmost was continuous over all others. The amount of this surface erosion has been greatest along

TABLE I

General Geologic Section for Minneapolis

Era	System and Series	Formation	Character of Strata	Approximate Maximum Thickness
Cenozoic	Pleistocene	Glacial drift	Bowlder clay, sand, gravel clay, alluvium, etc.	175 feet
Paleozoic	Ordovician	Decorah shale	Green limey shale	35
		Platteville limestone	Blue to gray thin-bedded limestone	25–35
		St. Peter sandstone	White or yellow fine-grained sandstone with some shale	175
		Shakopee dolomite	Yellow, buff, or pink dolomitic limestone	70
		New Richmond sandstone	White sandstone	40
		Oneota dolomite	Buff to reddish dolomite	100
	Cambrian	Jordan sandstone	Coarse-grained white sandstone	80–100
		St. Lawrence formation	Dolomite, shale, and a little sandstone	200 (?)
		Dresback sandstone	Fine-grained white sandstone, shaley toward base	250
Proterozoic	Algonkian	Red Clastic Series	Red sandstone and shale and some volcanic rocks	0— ?
Archean	Archean	Granite	Granite and gneiss	?

the courses of the old abandoned river channel and its tributaries. These ancient streams had cut entirely through the Decorah shale, Platteville limestone, and well into the St. Peter sandstone throughout their courses, and, as explained above, the trunk channel was so deeply eroded as to expose the Oneota in its bed along the lower portion of its course through the city.

This erosion was not confined to the pre-glacial streams but was, in a smaller degree, effective over the entire area. The Decorah shale, which is a persistent formation in Goodhue, Olmsted, Rice and other counties in southern Minnesota, has been almost completely eroded from the area included in the map. Only small isolated areas remain, capping some of the higher points in the south central and southeastern parts of the city and especially along the banks of the present channel of the Mississippi from the University campus southward to Minnehaha Creek.

Not only has pre-glacial and glacial erosion removed most of the shale from the area, but the limestone is also more or less eroded at all points within the city and completely removed from much of the northern part of Minneapolis, especially the northwestern portion along the area drained by Shingle Creek. Nevertheless, the Platteville limestone is the principal rock formation immediately below the drift and covers by far the greater part of the area within Minneapolis. This limestone is absent only along the stream channels, in the northwestern part of the area, and in a few scattered areas in north Minneapolis where erosion has reached the underlying sandstone. The distribution and relations of the formations within the city of Minneapolis may be seen on the map and sections accompanying this paper.

III. THE SURFACE OF THE BURIED ROCK

It will be seen from an inspection of the sections (Figs. 2 and 3) that the rock surface beneath the drift is generally nearly flat. The only notable features of relief are found along the old river channels which have been eroded to an average depth of 200 to 250 feet below the rock surface of the intervening upland plain. There are, however, numerous minor irregularities of surface due to the unequal erosion by the ice. The dip of the rocks is so flat as not to be noticeable to the eye except along the north edge of the city, where there is a low dip. Owing to the scarcity of outcrops this dip is only evident when the elevations of the top of the rock are compared to those in the central and southern part of the city. The top of the limestone along the south edge of Minneapolis stands at an elevation of about 790 to 800 feet above sea-level,

depending upon the amount of erosion which has occurred. In the extreme north edge of the city, in an old quarry about 3,400 feet east of the river, the top of the limestone is about 900 feet above the sea. About a quarter of a mile east of this quarry, underlying the Columbia Heights district, the limestone surface reaches its maximum elevation for the area, about 920 feet. This shows a rise of from 100 to 130 feet in a distance of about $8\frac{1}{2}$ miles. Most of the rise, however, is in the northernmost two miles, where the

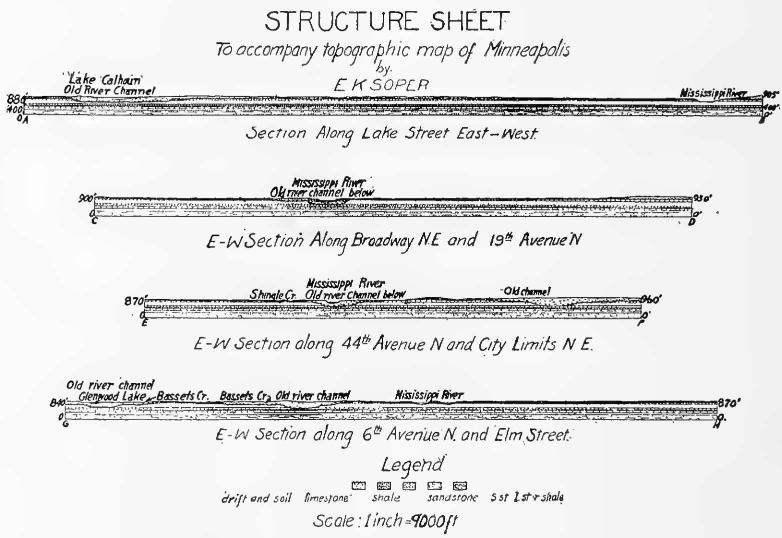

FIG. 2

dip steepens appreciably. The dip in the south half of the city is practically negligible. The elevation of the sandstone is correspondingly higher in the northern portion of the area, although the rise is not quite as great as in the case of the limestone, owing to the fact that the limestone is several feet thicker in the Columbia Heights district than in the south central portion of the city, where erosion has removed the upper layers. There are several minor undulations in the strata between the north and south city limits, but these are so slight as to be unimportant.

The generally flat surface of this limestone plain, back from the buried river gorges, is broken here and there by small hillocks.

These may be capped by a few feet of Decorah shale, which represent remnants of erosion. There are also several depressions or basins, probably gouged out by ice scouring, which, in several instances, are deep enough to extend through the limestone into the underlying sandstone. Another explanation for these basins is that they represent sink holes caused by the caving of the limestone over solution cavities produced in the underlying strata by ground waters. There is one cave about half a block long at Second Avenue South, between Fourth and Fifth streets, in the downtown district. This has not resulted in any noticeable surface depression, but in order to avoid such a possibility in the future, which might prove disastrous to buildings in the vicinity, it has been thoroughly walled up with concrete supports. This cave was due to mechanical action rather than to solution. The cavity occurs at the contact between the limestone and sandstone, but lies entirely within the sandstone. It is clearly due to the action of the ground water which has washed away the somewhat loosely consolidated sand at the contact, where the ground water tends to accumulate beneath an impervious bed of shale which separates the sandstone from the overlying limestone. This locality is only a few blocks from the river gorge along the sides of which the ground water comes to the surface. The production

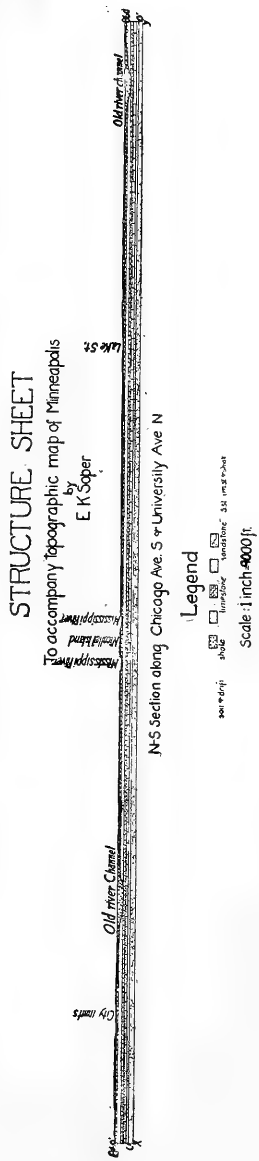

FIG. 3

of open cavities or channels in the sandstone by circulating ground

water was also observed in driving the tunnels through the sand rock on the University campus for the new heating plant. In the course of this work a fissure in the sandstone was encountered through which a large stream of water issued and temporarily flooded the tunnel. With this indisputable evidence of the occurrence of cavities in the rock below the surface, it is entirely possible that certain of the depressions in the limestone surface, as shown by excavations through the drift, may be due to the slumping of the limestone over similar cavities.

IV. RELATIONS OF THE BURIED ROCK SURFACE TO THE EXISTING DRIFT SURFACE

The major features of the present topography of the surface of the ground throughout the city of Minneapolis are in general independent of the underlying rock, and this is what one might expect in a region so heavily glaciated. There are, however, certain relations between the present surface and the buried topography which are suggestive and which, when understood, can be used to advantage in interpreting the character of the buried surface. The localities where there is an evident relation between the two surfaces are along the buried river channels. These old stream courses, which have been filled with débris to a depth of 200 to 250 feet, may still be roughly traced across the city by a series of depressions or low areas in the drift, some of which are now the sites of lakes. These depressions do not form a continuous and unbroken course in the present drift, for occasionally the independence of the glacial topography is manifested by abrupt morainal hills extending across the chain of depressed areas. An inspection of the accompanying map will show that nearly every natural lake within the city lies in the course of some buried stream. Not only are these pre-glacial channels indicated by a chain of lakes and shallow basins, but the general depression of the drift over the old stream beds has, at several places, resulted in determining the direction of the present surface drainage. Bassett's Creek and Minnehaha Creek both flow, for at least parts of their courses, directly over much larger drift-filled channels (see Fig. 1). Even the present channel of the Mississippi River, north of the mouth of Basset's Creek, to the city limits

and beyond, is superimposed upon the buried gorge of the preglacial river. Over the greater portion of the city, however, there is no suggestion in the surface topography of the irregularities in the buried rock surface. The topography of the existing surface is typically glacial in its character, and, although the maximum relief is not as great as in the case of the underlying rock surface, yet the general relief is seen to be stronger than that of the underlying rock surface, if we do not consider the old buried gorges. The lowest point on the surface within the city is at the bottom of the Mississippi River just below the mouth of Minnehaha Creek, in the extreme southeastern corner of the city, where the elevation is about 690 feet above sea-level. The highest points are along the ridge of morainal hills in the northeastern part of the city from Columbia Heights to a point about one-fourth mile southeast of Hillside Cemetery, where there are several points which attain an altitude of 965 feet, or thereabouts. This gives a maximum surface relief of about 275 feet. The lowest point accurately known on the surface of the bedrock was determined from the record of a well at the Bath House on the north shore of Lake Calhoun where the surface of the rock was encountered at an elevation of 619 feet. The highest point attained by the rock is in the Columbia Heights district where the top of the shale stands at about 930 feet. This gives a maximum relief of 311 feet for the rock surface, which is 36 feet greater than the relief of the existing drift surface.

There are probably two factors which are mainly responsible for the superimposed drainage system of this region and the general occurrence of low areas along the courses of the buried river valleys. Perhaps the most important one is the great depth to which these old gorges had been eroded. During the process of filling these valleys with glacial débris it may be that the amount of material available was not always quite equal to the amount required to fill them level with the tops of the banks at all points. This would result in slight depressions at intervals along the courses of the former streams. Another important factor is probably the settling of this newly deposited and loosely consolidated material. This settling would result from the saturation of the mass by ground water or from the gradual packing of the filling under its own

weight. In either case the shrinkage in a mass 200 to 250 feet thick might be sufficient to produce low undrained areas on the surface. Where these areas of depression are shallow and unimportant, settling could account for their existence. Where the surface basins are large and deep, as in the areas occupied by Lake Calhoun, Lake Harriet, Lake Nokomis, and that portion of the Mississippi River gorge north of the mouth of Bassett's Creek, their origin may be due to the incomplete filling of the pre-glacial gorge.

Because of the relatively even erosion over the flat areas between the pre-glacial valleys, there are few prominent peaks or high points of rock beneath the drift, and hence few outcrops are to be found in the city. The principal rock exposures are found along the post-glacial valley of the Mississippi from St. Anthony Falls to Fort Snelling and along the gorge of Minnehaha Creek, which is also post-glacial in age, from its mouth to Minnehaha Falls. Also at a few localities in North Minneapolis, where the drift is relatively thin, post-glacial erosion or artificial excavation has exposed the bed rock. There are also a few exposures near water-level on Nicollet Island and northward along the river to the mouth of Bassett's Creek.

V. THE BURIED RIVER CHANNELS

The history of the buried river channels of Minneapolis and vicinity, with an account of the origin and recession of St. Anthony Falls, has been worked out by N. H. Winchell,[1] U. S. Grant,[2] F. W. Sardeson,[3] and others. The general course of the main pre-glacial channel of the river across Minneapolis has been known for many years, but there has been no attempt to determine accurately the boundaries of the river and its tributaries and to construct a topographic map of the old rock surface beneath the drift that would

[1] *Geol. Survey of Minn., Fifth Annual Report* (1877), 175; also *Final Report*, II (1888), 313.

[2] "An Account of a Deserted Gorge of the Mississippi River near Minnehaha Falls," *American Geologist*, V (1890), 1.

[3] "The Beginning and Recession of St. Anthony Falls," *Bull. Geol. Society of America*, XIX (1908), 29–52.

show the depth and width of these buried gorges throughout their courses. The lack of data in former years made it impossible to construct such a map with any degree of accuracy. During the last decade, however, much additional data have been made available, especially at critical localities, through the sinking of wells and the construction of sewer tunnels and ditches and other excavations, so that it has become possible to show in some detail the character of the old drainage system. As will be seen from a study of the topographic map, the limestone which immediately underlies the drift over most of the city has been cut into six segments by pre-glacial stream erosion, and had St. Anthony Falls cut its way about one mile farther up the Mississippi to the mouth of Bassett's Creek before its progress was arrested by artificial means,[1] there would have been seven separate areas of limestone instead of six.

The present river follows the ancient course, which it has partially re-excavated, from a point beyond the north city limits to the mouth of Bassett's Creek in the north central part of the city. At this point the river leaves its pre-glacial channel, turns to the east, and flows southeastward through a young narrow gorge to Fort Snelling, at which point it is joined by the Minnesota River which flows into it from the south. As the map shows, the ancient river valley continued southward from the point where Bassett's Creek now has its outlet and flowed in a broad valley with gradually sloping banks southward across the region now occupied by the chain of lakes which comprises Lake of the Isles, Lake Calhoun, and Lake Harriet. At Bryn Mawr Meadows a tributary stream joined the river from the northwest, and its buried channel is now the course of Bassett's Creek between Glenwood Lake and Bryn Mawr Meadows. Glenwood Lake also lies within this tributary valley. A second tributary stream joined the main river from the west, just opposite Lake of the Isles. The presence of a buried valley east of this point is suggested by a series of basin-like depressions, one of which is the site of Cedar Lake. The existence of a third tributary valley opposite the south end of Lake Calhoun is indicated by well

[1] A protecting apron of timber was built over the rock at the point of fall to protect the rock from further erosion, which threatened to destroy the falls.

records. This valley also joins the main valley from the west, but there is little evidence of its existence in the present surface topography, which consists of low hills of drift and completely obliterates any underlying channel.

The pre-glacial river flowed almost due south from Lake Harriet, but no attempt has been made to outline its exact course beyond the city limits. In a general way this course can be traced for about five miles south of Minneapolis through Grass Lake, Wood Lake, and numerous small lakes and ponds to its junction with the Minnesota River valley somewhere east of the village of Bloomington.

An important tributary valley had its confluence with the main valley just south of Lake Harriet, at which point it entered from the east. Evidence of this buried valley may be seen in the present topography, especially along Minnehaha Creek, which follows the course of the ancient channel, between Lake Harriet and Lake Nokomis. At the latter point the old valley turns sharply to the south.

At Powderhorn Lake a small valley heads and flows southward to Rice Lake where it joins the tributary just described. This region around Powderhorn Lake is particularly interesting in this connection, for it has usually been considered to be a large Kettle hole.

North of the mouth of Bassett's Creek the Mississippi River flows over the same course that was formerly occupied in pre-glacial time. The old channel has been almost filled with sediments and drift so that the present stream bed, which consists of glacial débris and recent clays and silts, lies in this region at an elevation of at least 160 feet above the bottom of the buried valley. At a point about half way between the mouth of Bassett's Creek and the north city limits, along the river between Twenty-sixth Avenue North and Thirty-first Avenue North, a large tributary valley joined the pre-glacial river. This buried valley follows a course indicated in the present topography by a low swampy zone extending northeastward from the river through Sandy Lake (now dry) across Columbia Park, and beyond the city to Silver Lake. The exact course of the valley north of Silver Lake has not been determined.

The character of the material filling the ancient gorge in the north part of the city between the mouth of Bassett's Creek and the north city limits is strikingly different from that which constitutes the valley filling at all other points in the course of the main gorge and its tributaries. Along this particular portion of the river course the old valley is filled with fine laminated clays and silts instead of the heterogeneous drift which characterizes the filling at other places. The upper portions of these clay beds have been used for many years in the manufacture of brick and tile. These fine silts seem to extend from the bed of the present river (which at this place flows directly over the pre-glacial gorge) down to the bottom of the buried valley, a distance of about 160 feet. These silts also are found back from the main channel, where they cover many acres along the flats on both sides of the river. Their thickness is much less, however, away from the river.

This sudden change in the nature of the material filling the old valley, the suggestive laminated structure of the upper portion of the deposit, the variation in thickness of the material, and its occurrence back from the river channel, along the flood plain of the river, are all evidences which point to an original deposition in a lake. The change from gravel and drift to clay and silt occurs immediately above the point where the present river abandons the pre-glacial course. It seems probable that at a late stage in the recession of the ice a moraine or dam was formed across the old gorge near the mouth of Bassett's Creek. This resulted in the formation of a temporary lake which filled the river valley, overflowed the banks, and spread over the low land which flanked the valley on both sides. In this lake the fine clays and silts could be deposited. Eventually an outlet was formed for the impounded waters at the point where the present river diverges from the filled gorge. After the outlet was once established the erosion of the new post-glacial gorge was started. But deposition of fine sediments could continue for a long time after the new gorge was started, because the lake would still remain and the current would continue to be slackened to the point where clays and slits could accumulate. The boundaries of this lake have not been traced north of the city limits, but it is probable that it extended a considerable distance beyond them.

VI. THE DRIFT

The glacial drift which was spread over the land upon the retreat of the ice, and which covers the rock surface throughout the city, is variable in its composition. It ranges from typical bowlder clay, which consists of a heterogeneous mixture of clay, sand, gravel, and bowlders, to sorted deposits of fine sand or clayey loam. Most of the drift consists of a mixture of clay, sand, and bowlders of all sizes. Where the sand in the drift is free from clay and gravel, it may have been deposited in two ways: either it represents a wind-blown deposit and originated as sand dunes, or it is a flood-plain deposit, washed up by glacial streams and sorted by water action. The only place in the city where clay occurs in large quantity free from pebbles or bowlders is along the banks of the Mississippi River north of the mouth of Shingle Creek. These clay beds probably represent lake and river deposits rather than accumulations due to glacial action. They are indirectly the product of glacial action, since they were deposited during the Glacial stage of the river.

Since the surface of the drift constitutes the surface of the land, except where it has been covered by post-glacial deposits of alluvium, sand, or *loess* of slight thickness, it controls the topography of the region. Therefore, what has been said concerning the characteristics of the topography of the city will apply also to the drift.

The thickness of the drift is variable and ranges from 0 to 250 feet. As shown by the scarcity of outcrops (see map) the drift is present nearly everywhere except along the river bluffs. The localities where it is thinnest are those which border the rock outcrops. In north Minneapolis there is an area bounded by Central Avenue on the south and east, Twenty-third Avenue on the north, and Main Street on the west, where the drift is thin, ranging from 0 to 15 or 20 feet. Throughout the greater part of South Minneapolis, in the large area bounded by the Mississippi River on the east and north, the pre-glacial Mississippi River valley on the west, and Minnehaha Creek on the south, the thickness of the drift is very uniform. The average depth over most of this area is from 30 to 50 feet, the greatest thickness being in the southern portion. Both the surface of the drift and the surface of the underlying limestone are generally flat over this region. The only noteworthy breaks in

the otherwise almost flat surface are at Powderhorn Lake, where a small buried valley heads, and at Lowry Hill and the region around Loring Park, where a belt of low morainal hills extends from the west across the edge of the area and breaks the monotony of the surface.

In some portions of the city, especially in Southeast Minneapolis, in the vicinity of the University and northeastward to the city limits, there is a bed of peat several feet thick which occurs only a few feet below the surface. This peat bed rests upon drift largely composed of sand, clay, or gravel, or a mixture of the three, and is covered by recent deposits of sand or soil which are of post-glacial or glacial origin.

In East Minneapolis, near Tower Hill in the Prospect Park district, deposits of *loess*, a formation of loamy material of aeolian origin, occur at the surface. Therefore, it is apparent that while the retreating ice left a mantle of drift over the entire region, the surface of this drift has been slightly modified by post-glacial deposition. These deposits of sand, silt, and other material were chiefly laid down as flood-plain deposits in the streams and marshes fed by waters from the melting ice. A small amount of post-glacial erosion has also operated to modify the original drift surface, but this has been unimportant except along the channel of the river and its tributaries.

The origin, composition, and character of different drift deposits and their modification by post-glacial agencies have all been described in detail in reports by N. H. Winchell,[1] Warren Upham,[2] F. W. Sardeson,[3] and F. F. Grout and E. K. Soper.[4]

VII. INFLUENCE OF BURIED RIVER CHANNELS UPON BUILDING CONSTRUCTION

In those portions of the city which are underlain by the buried valleys and in other parts of the city where the drift is deep, it is impracticable to carry excavations to bedrock for building foundations and cellars. For the ordinary small store or residence, or for

[1] *Final Report Minn. Geol. and Nat. Hist. Survey.* [2] *Ibid.*

[3] "Geology of the Twin Cities," *Minneapolis and St. Paul Folio, U.S. Geol. Survey.* Now in press (1914).

[4] Grout and Soper, "The Clays and Shales of Minnesota," *Bull. No. 11, Minnesota Geological Survey,* chapter on Hennepin County (1914).

light buildings of any description, it is not necessary to have a perfectly rigid foundation. The drift throughout Minneapolis is usually fairly solid at moderate depths below the surface and is quite safe for foundations for buildings of average size. But in the regions along the courses of the buried river valleys, where the drift may be from 100 to 250 feet deep, areas of loose sand which often has a tendency to run are frequently encountered. In such localities it has been found necessary to use piling in constructing foundations for large buildings.

The main business section of the city lies back from the buried valley and in the downtown district the average distance to bed-rock is only 25 to 30 feet. The rock in this locality consists of lime-stone or hard shale and can easily be reached if necessary in building operations. For this reason it has been necessary to use piling in the construction of only a few buildings in the city. Should the zone of large buildings north of Hennepin Avenue be extended in the future a few blocks farther west beyond the Great Northern Railroad tracks, it will be found necessary to use piling, caissons, or some type of foundation adapted to soft ground, for many if not all of the large structures. The side of the buried valley in this vicinity drops off rapidly, so that points along the present channel of Bassett's Creek overlie the very bottom of the old valley nearly 200 feet below. Much of the material filling the old gorge in this vicinity is loose sand. At all points in the city, except those along the sides and courses of the buried valleys, the foundations are generally good.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA FOR 1909, 1910, 1911, AND PART OF 1912

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IV. ONTARIO, NORTHWEST TERRITORIES, AND THE EAST COAST OF HUDSON BAY

Allen[1] states that the Woman River area in the Sudbury Mining Division of Ontario, west of Rush Lake, shows the following succession:

Basic igneous dikes } Relative ages not known
Mica porphyry }
Acid igneous rocks—extrusive and intrusive
Iron formation
Basal greenstones

The series has been intensely folded and extensively brecciated. The strike of the iron formation banding is N. 45° E., but is in some places at right angles to this direction where modified by cross-folding. Everywhere the rocks stand practically on edge. The basal greenstones and the iron formation are intruded by a great many basic and acid dikes, the latter showing gradation into an acid volcanic breccia overlying the iron formation.

The various phases of the iron formation are: (1) finely banded cherty carbonates, (2) hematitic, magnetic, pyritic, cherts, (3) black and red jaspilites, (4) a unique amphibole-magnetite rock, in which the petrographic character of the amphibole suggests riebeckite, and (5) iron ore. The cherty iron carbonates are regarded as the source of the other iron-bearing rocks, although the possibility that some of the varieties may be partly original is not denied. The derivation of the secondary varieties is ascribed to the same katamorphic and anamorphic processes by which Van Hise has explained the origin of similar derivatives in the other iron formations.

[1] R. C. Allen, "Iron Formation of Woman River Area," *Eighteenth Annual Report of the Bureau of Mines*, 1909, Ontario.

The distribution of the iron ores bears no relation to the present structural features of the formation nor to its present erosion surface.

The relation of the iron formation to the associated igneous rocks is expressed by the following statements: The iron formation is sedimentary; it conformably overlies the ellipsoidal greenstones, in that no period of subaerial erosion intervened between the deposition of the iron formation and the greenstone; the iron formation and the overlying volcanic breccia and rhyolite porphyry are also conformable. The evidence of the sedimentary origin of the iron formation consists in its banding, the iron carbonate content, and the parallelism of the individual bands to each other and to the basal plane.

The evidence for conformity with the basal greenstone lies in the parallelism of the banding of the iron formation with the plane of contact with the greenstone, the absence of detrital material, and the probable subaqueous origin of the ellipsoidal greenstones. The conformity of the iron formation and the overlying acid igneous rocks appears probable but is not supported by many field observations, owing to the sparsity of exposed contacts. Wherever observed, the contacts are sharp and show no detrital materials.

From field relations of the iron formation to igneous rocks in the Woman River and other areas but mainly from the work of Van Hise and Leith, Allen believes that the genesis of the iron formations was related to igneous rather than sedimentary agencies and processes. He interprets the physical history of this area to be extrusion of basic greenstones upon a submerged floor, rapidly followed by the precipitation of iron formation, contributed from magmatic solutions, without the intervention of clastic sedimentation. The precipitation of the iron formation was followed by the extrusion of acid igneous rocks. The three rock types, however, basal greenstone, iron formations, and acid igneous rock, are perhaps differentiates of one parent-magma.

Adams[1] describes the occurrence of silver and calcite, in wall rock adjacent to a fissure vein of the Cobalt district.

[1] F. D. Adams, "Notes on the Occurrence of the Ore Body at the City of Cobalt Mine," *Jour. Can. Min. Inst.*, XII (1909), 414–17.

Baker[1] states that the pre-Cambrian rocks of the Lower Mattagami basin are pink gneisses of Laurentian age, Post-Middle Huronian diabase dikes, a small outcrop of a banded siderite, quartz conglomerate, and quartzite which may be Upper Huronian.

Baker[2] states that the pre-Cambrian in the Lake Abbitibi area consists of Keewatin schistose, ellipsoidal greenstones, and other metamorphosed eruptive rocks infolded with dolomites, graphitic slates, jaspilites, and fragmental materials, all intruded by Laurentian granites, pegmatites, and granite porphyries. This old complex is intruded by post-Huronian quartz diabase, quartz gabbro, and Camprophyric and aplitic dikes. The diabase is associated with veins showing traces of silver and gold, but no important metal deposits have thus far been found.

Baker[3] states that the pre-Cambrian rocks of the Mattagami River region are Laurentian granites and gneisses intruded by diabase dikes, probably Post Middle Huronian in age, Upper Huronian siderite, sideritic conglomerate, and quartzite.

Bancroft[4] reports the following pre-Cambrian formations from the Keekeek and Kewagama Lakes region, 40 miles northeast of Lake Temiskaming.

Post Huronian	—dikes and stocks of basic intrusives probably Keweenawan
Huronian (?)	—conglomerate, arkose, and graywacke—all schistose
Laurentian	—batholiths and stocks of granite and syenite
Keewatin	—basic and acid schists, iron formation, tuffs, volcanic breccias, rhyolites, basalts, basic and acid extrusives

Burrows[5] states that the South Lorrain silver area is north of the Montreal River and west of Lake Temiskaming, and about 16 miles southeast of the Cobalt district. The pre-Cambrian rocks are Keewatin, basic igneous rocks, and minor acid intrusives, intruded

[1] M. B. Baker, "Iron and Lignite in the Mattagami Basin," *Ont. Bur. Mines, 20th Ann. Rept.*, pp. 214–46, 37 figs.

[2] M. B. Baker, "Lake Abbitibi Area," *Ont. Bur. of Mines, 18th Ann. Rept.*, XVIII (1909), 263–83, pl. 1, 9 figs.

[3] M. B. Baker, "The Iron Ores of the Mattagami River," *Jour. Can. Min. Inst.*, XIV, 299–309.

[4] J. Austin Bancroft, "Report on the Geology and Mineral Resources of Keekeek and Kewagama Lakes Region," *Province of Quebec*, 1911, pp. 160.

[5] "South Lorrain Silver Area," *Ont. Bur. Mines, 18th Ann. Rept. (1908) pl. 2*, pp. 21–31, 1909.

by Laurentian granite and syenite; Huronian quartzite, arkose, conglomerate, slate, and breccia, unconformably above the Keewatin and Laurentian; and post-Huronian diabase intrusives. Quartz and calcite veins bearing native silver, smaltite, and niceolite have been found near contacts between the post-Huronian diabase intrusives, and Keewatin basic igneous rocks.

Burrows[1] reports that the pre-Cambrian succession in the Gowganda and Miller Lake silver area between Sudbury and Cobalt is Keewatin altered basic igneous rocks and acid porphyries, intruded by Laurentian granite, syenite, and gneiss; unconformity; Huronian quartzite, arkose, graywacke, conglomerate, and slate; intrusive into all the preceding, post-Middle-Huronian diabase. The diabase contains silver-bearing veins whose value has not been determined.

Burrows[2] finds that the compact rocks of the Porcupine district 100 miles northwest of the Cobalt district, are all pre-Cambrian, principally Keewatin. Keewatin and Huronian rocks contain irregular gold-bearing veins, in which quartz is the dominant mineral.

Bowen[3] reports that sills and dikes of diabase cut older formations in the Gowganda Lake district. The dikes, some of them 250 feet wide, show no evidence of differentiation. Red spots composed of an intergrowth of quartz and soda-rich plagioclase feldspar occur in the diabase of the sills. At the contact of sills and slate, a granophyre consisting of soda-rich plagioclase, quartz, and : ccessory minerals including calcite grades into a slate adinole developed by contact metamorphism. The granophyre and adinole are both ascribed to hydrothermal actions, the granophyre being regarded as an extreme phase of the alteration of the slate, a recrystallization from a state of aqueous fusion.

To the reviewer, it seems that the red spots in the sills, and the granophyre, may be due to the same causes, since they are

[1] A. G. Burrows, "The Gowganda and Miller Lakes Silver Area," *Ont. Bur. Mines, 18th Ann. Rept.*, XVIII (1908), Pt. 2, pp. 1–20, 20 figs., 3 maps.

[2] A. G. Burrows, "Porcupine Gold Area," *Quart. Bull. Canadian Min. Inst.*, No. 16, 1911, pp. 59–62.

[3] Norman L. Bowen, "Diabase and Granophyre of the Gowganda Lake District, Ontario," *Jour. Geol.*, XVIII, No. 7 (1910), 658–74.

mineralogically similar and are gradational. Bowen believes that the granophyre may be quite different in origin from the red spots scattered through the sills, despite their similarity. A somewhat similar phenomenon in the Cobalt district, the separation of diabase and aplite, has been ascribed to differentiation by Collins and Hore, Collins showing that the separation depended on the rate of cooling, being entirely absent in diabase, which was rapidly chilled. A magmatic separation of soda-rich aplite and gabbro in the intrusive of Mt. Bohemia, Mich., has been described by F. E. Wright. Here there is no evidence that the aplite was associated with slate.

Bowen[1] presents chemical analyses and descriptions of the diabase and aplite of the Cobalt silver area.

Bowen[2] describes the rocks of the Thunder Bay district as Upper Huronian argillites, gray quartzite and gray slates interstratified, black slate, and iron formation, all unconformable above the Laurentian granite complex, and intruded by Keweenawan diabase dikes and sills. Silver veins are found in steeply dipping fault zones cutting the black slate in an east-west direction. The ore minerals are principally native silver and argentite, associated with calcite, quartz, fluorite, barite, witherite, and other minerals.

Coleman[3] states that the rocks of the Black Sturgeon Lake district southwest of Lake Nipigon are Keewatin green schists intruded by granite gneiss, and slightly disturbed Keweenawan diabase, shale, and sandstone. Hematite veins are found in fault zones of the Keewatin.

Coleman[4] reports that the Alexo nickel deposit near Matheson, northern Ontario, consists of a pyrrhotiferous nickel-bearing rock of unknown extent, grading upward into an ultra basic rock, peridotite altered to serpentine, and resting on footwall of andesite.

[1] N. L. Bowen, "Diabase and Aplite of the Cobalt-Silver Area," *Jour. Can. Min. Inst.*, XII (1909), 517-28.

[2] N. L. Bowen, "Silver in Thunder Bay District," *Ont. Bur. Mines, 20th Ann. Rept.*, pp. 119-32.

[3] A. P. Coleman, "The Black Sturgeon Lake District," *Ont. Bur. Mines, 18th Ann. Rept.*, Pt. 1, pp. 163-179.

[4] A. P. Coleman, "The Alexo Nickel Deposits," *Econ. Geol.*, V, No. 4 (1910), Pt. 4, 372-76.

Coleman interprets the deposit as a marginal magmatic segregation from the peridotite.

Collins[1] finds that the pre-Cambrian rocks in the region between Lake Nipigon west of Clay Lake, a distance of about 220 miles, include Keewatin and Huronian altered acid and basic intrusives, volcanic tuffs, chlorite, and sericite schist with bands of conglomerate, graywacke, phyllite, quartzite, and jaspilite; Laurentian granites, syenites, and diorites intrusive into the Keewatin; post-Keewatin eruptives, gabbro, mica syenite, acid porphyries, and granite; flat-lying Keweenawan impure dolomite and sandstone, and post-Keweenawan diabase intrusive sheets.

Collins[2] reports that the pre-Cambrian rocks between the Pic and Nipigon rivers of the north shore of Lake Superior include Laurentian granite, gneisses, and syenite; Keewatin graphitic gneiss, quartzite, arkose, amygdaloids, and basic and acid schists; Keweenawan red arenaceous dolomite, and diabase intrusives; eruptives of unknown age, viz., syenites, diorite, pegmatite, and diabase.

Collins[3] states that the cobalt silver ores of northern Ontario are associated with diabases. The diabasic magma has acted as a mixture of two rock species, diabase and aplite, whose segregation was controlled by the rate of cooling, no separation having taken place in rapidly cooled rocks. The aplites show differentiation into quartz veins, and still later quartz calcite veins. The diabase contains some chalcopyrite, the aplite still more, while an even higher percentage is contained in the quartz calcite veins. Neither silver nor cobalt were found in the diabases.

Harvie[4] states that the succession of Keewatin, Laurentian, and Huronian rocks of the Oposatica district, about 40 miles east of

[1] W. H. Collins, "A Geological Reconnaissance of the Region Traversed by the National Transcontinental Railway between Lake Nipigon and Clay Lake, Ont.," *Canada Geol. Surv. Branch*, 1909. 67 pp., 2 pls., 1 fig., 2 maps.

[2] W. H. Collins, "Report on the Region Lying North of Lake Superior betwen Pic and Nipigon Rivers, Ont.," *Canada Geol. Survey*, 1909, 24 pp., 1 map.

[3] W. H. Collins, "Quartz Diabases of Nipissing District," *Econ. Geol.*, V, No. 6 (1910), 538–50.

[4] Robert Harvie, "The Telluride Ores at Oposatica," *Jour. Can. Mining Inst.*, 1911, pp. 164–70.

north from the north end of Lake Temiskaming, is similar to that of the Cobalt district. Narrow gold-bearing quartz ankerite veins are found cutting Keewatin and Huronian rocks in the Oposatica district. The gold minerals are petzite and native gold. The petzite occurs in cracks in the quartz and ankerite, and the native gold was deposited after the petzite. Harvie believes that the Oposatica ores will show no increase in values with depth, since the Tellurides in known districts have all been found below the weathered zone.

Hore[1] discusses the stratigraphy and the ores of the Cobalt district. The richest ores have been found in the Huronian conglomerate. Rich shoots have also been mined in the Keewatin greenstone, and in the Keweenawan diabase. The ores are genetically connected with the diabase intrusions.

Hore[2] argues that the cobalt silver ores, and the aplites and diabases with which they are associated are all differentiation products of one magma.

Hore[3] believes that the glacial and glacio-fluvial origin of a portion of the conglomerate-quartzite-shale series of Huronian age at Cobalt and Temagami is clear. The evidence consists in the small amount of stratification, the presence of striated and soled pebbles, the heterogeneity of the conglomerate, and the discordance between the lithologic character of the conglomerate and the basement on which it was deposited. The basal conglomerate may not be of glacial origin since they are like the materials on which they lie. Some of the materials are probably water deposits, into which large erratics were introduced by floating ice. Certain thick bowlder-free deposits of shale and graywacke are probably ordinary water-laid sediments.

Hore[4] states that Keewatin and Huronian rocks similar to those of the Nipissing area outcrop in the Porcupine district. The

[1] R. E. Hore, "Geology of the Cobalt District," *Trans. Am. Inst. Min. Eng.*, XLII (1912), 481–98.

[2] Reginald E. Hore, "Differentiation Products in Quartz Diabase Masses of the Silver Fields of Nipissing, Ontario," *Econ. Geol.*, VI (1911), 51–59.

[3] Reginald E. Hore, "Glacial Origin of Huronian Rocks," *Jour. Geol.*, XVIII, No. 5 (1910), pp. 459–67.

[4] Reginald E. Hore, "On the Nature of Some Porcupine Gold Quartz Deposits," *Jour. Can. Min. Inst.*, XVI, No. 15, pp. 57–70.

Keewatin rocks include acid and basic extrusives, some plutonics jaspilitic iron formation, ferro-dolomites, and schists of unknown origin. The Huronian rocks are chiefly quartzite, graywacke, micaceous schists, and one conglomerate, apparently made up of igneous materials.

Most of the gold ores occur in quartz-bearing veins within metamorphosed igneous rocks of the Keewatin or the Huronian conglomerate. The quartz is full of liquid inclusions and in places is shattered. Gold occurs between quartz veins, included in quartz, in pyrite, and in calcite.

Hore believes that the ores and quartz pyrite gangue were deposited by hot alkaline solutions. He finds that they resemble the ores of the mother-lode closely and that the same hypothesis of origin is applicable to both.

The pre-Cambrian succession in the Nipissing district according to Hore[1] is:

I. Algonkian.
- *a*) Keweenawan. Igneous intrusions of quartz diabase, quartz gabbro with acid differentiation products. Some olivine diabase and diabase porphyrite dikes. Igneous contact.
- *b*) Huronian.
  - (1) An upper series, probably Middle Huronian. Chiefly feldspathic quartzite with some conglomerate. Barlow's Lorain series.
    Slight unconformity.
  - (2) A lower series, probably Lower Huronian, graywacke, shale, conglomerate, and feldspathic quartzite. Miller's Cobalt series.
    Great unconformity.

II. Archaean.
- *a*) Laurentian. Granites, diorites, syenites, gneisses intrusive into Keewatin.
- *b*) Keewatin. Basic and acid extrusives and intrusives, felsite agglomerate, iron formation, slates and green schist.

The silver ores occur principally in narrow vertical or nearly vertical fissure veins cutting Keweenawan diabase, the lower Huronian series, and the Keewatin. The veins strike in various directions, but are uniform in composition, consisting mainly of smaltite, calcite, and silver, deposited in the order in which they

[1] "The Silver Fields of Nipissing, Ontario," *Can. Min. Inst. Quart. Bull.*, No. 17 (1911), pp. 81–105, 3 pls.

are named. The richest veins are in the Huronian. Hore supports the views that the silver-bearing solutions came from the Keweenawan diabase, that the Keewatin greenstone played an important rôle in their deposition, and that the coarse Huronian sediments afforded abundant openings for their reception.

Hore[1] states that the diabase of the Cobalt district occurs in dikes and sills intrusive into Keewatin schists, and Huronian sediments. The minerals of the diabase are pyroxenes, plagioclase, quartz ilmenite, biotite, and olivine. Olivine seems to be confined to small masses of diabase showing no differentiation. Aplitic micropegmatitic intergrowths of sodic feldspar and quartz, sometimes with calcite, occur in the thick sills. Some of the micropegmatite is clearly a differentiate of the diabase *in situ*, occurring in spots and stringers grading into the diabase. Some of it is a later differentiate, filling fissures sometimes with selvage borders.

A minor fissure filling in the diabase consists of quartz, with chlorite and amphibole. Galena, pyrite, and chalcopyrite are frequently present. Calcite sometimes fills interstices between well-formed crystals of calcite.

King[2] describes a lens-shaped deposit of low phosphorus, non-titaniferous magnetite ore, which occurs at the contact between Grenville limestone and Laurentian granite at the Wilbur mine, in Lanark County of eastern Ontario.

Knight[3] presents diagrams showing faults in the Cobalt district. The largest known fault has a displacement of over two hundred feet. It is of the reversed type and strikes parallel to the longer axis of Cobalt Lake.

Kerr[4] presents a petrographic description of the Port Caldwell syenite area on the north shore of Lake Superior about 125 miles east of Port Arthur. The principal rock of the region is nepheline

[1] Reginald E. Hore, "Diabase of the Cobalt District, Ontario," *Jour. Geol.*, XVIII, No. 3 (1910), 270f.

[2] Shirley King, "The Wilbur Iron Mine," *Jour. Can. Min. Inst.*, XII (1909), 582–91.

[3] C. W. Knight, "Recent Underground Development Work at Cobalt," *Jour. Can. Min. Inst.*, XV, No. 18 (1912), pp. 21–23, 1 pl.

[4] H. L. Kerr, "Nepheline Syenites of Port Caldwell," *Ont. Bur. Mines, 19th Ann. Rept.*, Pt. 1, pp. 194–232, 20 figs., 1 map.

syenite with small amounts of hornblende, augite, and magnetite. Several other syenite facies are represented, besides pegmatites, small bodies of gabbro, granite, diorite, and later basic dikes.

Lane[1] presents a detailed description of the records of 17 drill holes, placed along a line normal to the strike of the Keweenawan copper-bearing rocks on Point Mamainse on the east shore of Lake Superior. The copper-bearing series in this cross-section consist of roughly parallel beds of conglomerate, aplite, and porphyrite intruded by felsite. Their strike is about N. 10° W. and their dip is from 20°–50° west or lakeward. The beds are frequently offset by dip faults.

Leith[2] states that the Algonkian, Nastapoka, and Richmond groups, exposed on the east coast of Hudson Bay, consist of slightly metamorphosed chemical and mechanical sediments, interbedded with basic eruptives, with a seaward dip from 50° to 45°. The Richmond group, unconformably, below the Nastapoka, embraces coarse, ill-assorted mechanical sediments interbedded with basic eruptives, generally dipping seaward at steeper angles than the Nastapoka. The Nastapoka contains an iron formation containing greenalite, iron carbonate, and derivative phases closely related to extrusives. The Nastapoka is probably late Algonkian, perhaps Animikie, because of its structural, stratigraphic, and metamorphic similarity to the Animikie of the Lake Superior region. Leith suggests that during the Algonkian, the Archaean protaxis may have separated the Lake Superior geo-syncline of deposition on the south from a similar region of deposition on the north.

Mickle[3] finds that one hundred square miles of the Nipissing region have numerous calcite veins, but that the metal-bearing veins are restricted to ten square miles. On these ten square miles there are about 2,000 calcite veins of which about 1 per cent are metal-bearing, and only 66 veins were productive at the end of 1907. About 24 per cent of the area is underlaid by Huronian

[1] A. C. Lane, "Diamond Drilling at Point Mamainse, Province of Ontario," *Canada Dept. of Mines, Bull. No. 6,* 1911, 59 pp., 1 fig., 1 map, 5 pls.

[2] C. K. Leith, "An Algonkian Basin in Hudson Bay," *Econ. Geol.*, V, No. 3, pp. 227–46.

[3] G. R. Mickle, "The Probable Number of Productive Veins in the Cobalt District," *Jour. Can. Min. Inst.*, XIII (1910), 325–35.

rocks, 42 per cent diabase, and 33 per cent by the Keewatin. The Huronian had 53 productive veins, the Keewatin 6, and the diabase 7. Assuming that these areas have prospected equally well, the chances of finding ore are about 9 times better on the Huronian than on an equal area of Keewatin, and about one-twelfth as good on the diabase as on an equal area of Huronian.

Miller[1] reports that the pre-Cambrian rocks exposed in the district of Patricia, comprising over 146,400 square miles west of James Bay, which has recently been added to the Province of Ontario, include limestones similar to those of the Hastings and Grenville districts, a small volume of Lower-Middle Huronian conglomerates, iron-bearing Animikie, and diabase intrusives, probably Keweenawan.

Miller and Knight[2] state that they have found an unconformity between the Grenville and the Hastings series of eastern Ontario, the Hastings being younger. The Grenville, they maintain, has been deposited on rocks similar to the Keewatin greenstones of Lake Superior, and the Grenville, therefore, may be the time equivalent of the Keewatin iron formations farther west. The Hastings series they regard as Huronian.

Moore[3] describes the geology of the Sturgeon Lake goldfield in 1911. It is located in latitude 50° north longitude, northwest of Thunder Bay. The rocks are pre-Cambrian, largely igneous, including granites, gneisses, aplites, quartz-porphyries, rhyolites, hornblende-syenite, diorites, diabase, basalts, gabbros, and porphyrites and their metamorphic equivalents. There are small patches of arkose, graywacke, and dolomite. The relations of the igneous rocks to each other is partially known, but the age of the sediments and the igneous rocks and their relations are still uncertain. Free-milling gold veins are found at contacts, and in cleavage and fault planes and other openings. The gangue is principally

[1] W. G. Miller, "District of Patricia," *Ont. Bur. Mines,* XXI, Pt. 2, pp. 200, 18 figs., 29 maps.

[2] W. G. Miller and C. W. Knight, "Grenville-Hastings Unconformity," Abstract, *Geol. Soc. Am. Bull.,* XIX (1909), 539–40.

[3] E. S. Moore, "The Sturgeon Lake Gold Field," *Ont. Bur. Mines, 20th Ann. Rept.* pp. 133–57, 17 figs.

quartz, with small amounts of calcite and siderite. Sulphides of copper, lead, zinc, and iron are also present.

Moore[1] states that the rocks at Round Lake north of Lake Nipigon are Keewatin ellipsoidal greenstone, green schist, fine-grained quartose gneiss, and lean, magnetic, cherty iron formation, intruded by Laurentian granite and gneiss. The Archaean complex is intruded by Keweenawan diabase. The iron formation is separated from the greenstones by schists and sediments. It consists of bands of crystallized chert, magnetite, hematite, and some siderite.

Moore[2] states that the pre-Cambrian rocks of the Lake Savant Area northwest of Lake Nipigon belong to the Keewatin, Laurentian, Huronian, and possibly Keweenawan systems. The Keewatin comprises acid, intermediate, and basic extrusives and intrusives, banded magnetic, cherty iron formation, and an unusual development of sediments in the form of graywacke and gray gneiss. The Laurentian consists of granites and gneisses intrusive into the Keewatin. The rocks of the Huronian are principally conglomerates, graywacke, and a little quartzite. A few basic dikes may be of Keweenawan age.

Moore[3] describes the pyrite deposits of Big Vermilion Lake near the Minnesota boundary. The pre-Cambrian rocks of this region include a basement of intrusive and extrusive acid and basic rocks, overlaid by cherty iron formation, graphitic slate, graywacke, arkose and quartzite. The two systems were intruded by granite. The pyrite occurs in fissure veins in greenstone, quartz porphyry, and iron formation.

Moore[4] states that the Onaman Iron Range district covers about 70 square miles and lies between 45 and 50 miles up the Red Paint River, northeast of Lake Nipigon. The Keewatin, Laurentian,

[1] E. S. Moore, "Iron Range North of Round Lake," *Ont. Bur. of Mines, 18th Ann. Rept.*, XVIII, Pt. 1, pp. 154–62.

[2] E. S. Moore, "Lake Savant Iron Range," *Ont. Bur. Mines, 19th Ann. Rept.*, XIX (1910), Pt. 1, pp. 173–93, 16 figs.

[3] E. S. Moore, "Vermilion Lake Pyrite Deposits," *Ont. Bur. Mines, 20th Ann. Rept.*, 1911, pp. 199–208, figs. 6, 1 map.

[4] E. S. Moore, "Geology of Onaman Iron Ranges Area," *Ont. Bur. of Mines, Ann. Rept.*, XVIII (1909), Pt. 1, pp. 196–253, 22 figs., 3 pls., 1 map.

Huronian, Keweenawan, Pleistocene, and Recent rock systems are represented. The Keewatin consists of greenstone and greenstone schists, rhyolite, rhyolite porphyry, and feldspar porphyry, rhyolite tuff, agglomerate, and conglomerate, and iron formation. The Laurentian consists of granite and granite-gneiss, batholiths, intrusives into the Keewatin. The Keweenawan and Huronian are represented by intrusives. Pleistocene drift and Recent alluvial and travertine deposits cover a considerable portion of the district.

The Keewatin iron formation occurs in two nearly parallel, synclinorial belts running nearly east and west. Folding was accomplished by fracture and flow, predominantly the latter, although locally the formation is intensely broken and even faulted.

The various iron formation phases consist of cherty iron carbonate, ferruginous cherts, actinolite magnetite quartz rocks, interbedded with clastics whose grain suggests that they are of pyroclastic origin. The cherty iron carbonate and some of the iron oxide phases are probably original. The alterations of the iron formation have been chiefly anamorphic. Among the secondary minerals is dumortierite, a unique basic aluminium silicate, never before reported from iron formation.

Moore finds difficulty in accounting for the clean-cut banding of the iron formation as well as its sharp contact with the interbedded clastics on the hypothesis that the iron salts were contributed by the weathering of basic igneous rocks. These characters he finds in perfect accord with Leith's theory of direct igneous contribution. However, direct evidence for the latter he believes is lacking, since the iron formation is nowhere in direct contact with basic igneous rocks excepting at faults, and is also separated from rhyolite by slates, and what appear to be water-deposited pyroclastics, containing no iron formation. If the clastics in association with the iron formation are pyroclastic in origin, considerable difficulty in the way of the direct-contribution hypothesis would be removed. The iron and silica could have been delivered as well by weathering as by direct contribution in Moore's opinion. He cites the solubility of the silica in the iron formation as shown by Van Hise and Leith, the silica content of bog ores described by Hunt and himself as evidence of the adequacy of weathering in causing the deposition of silica.

While the evidence for the solubility of silica under weathering may be satisfactory, there is very little to show that weathering may yield siliceous deposits of anything like the thickness, high silica content, and freedom from detrital silica and clay as the Lake Superior sedimentary iron formations.

Moore believes that the Keewatin was a period of inclosed basins and imperfect drainage, conditions especially favorable to the solution, transportation, and deposition of iron. The long period of freedom from intense vulcanism, which he believes is implied by the thickness of the iron formation, removes the most apparent difficulty in assuming the development of a luxuriant flora, a condition conducive to the solution of iron and silica, and a selective retention of the complementary clastics on the land areas. Unless it can be shown that the clastics associated with the iron formation are pyroclastic in origin, it is obvious that a great deal of rock was disintegrated by weathering during the Keewatin. With due regard for the facts which may be interpreted as evidence for weathering and magmatic contribution, Moore concludes that the materials of the iron formation were supplied to inclosed basins by weathering action under the influence of plant life, and by heated igneous rocks coming in contact with the waters.

McInnes[1] reports that the pre-Cambrian rocks of the region between N. lat. 50° 10′ and N. lat. 55° 10′, and between W. long. 86° and W. long. 90°, include Keewatin diorites, diabases, and chloritic and hornblende schists, and coarse conglomerates, the latter probably Huronian, and Laurentian biotite granite gneisses.

Parsons[2] writes detailed descriptions of the gold prospects and mines of the Lake of the Woods, Manitou, and Dryden areas of northwestern Ontario.

Seelye[3] states that the Helen limonite ore body at Michipicoten is bounded by cherty iron carbonate, chert, and a diabase dike.

[1] W. McInnes, "Report on a Part of the Northwestern Territories Drained by the Umisk and Ottawapiskat Rivers," *Canada Geol. Survey*, 1910, 54 pp., 5 pls., 1 map.

[2] A. L. Parsons, "Gold Fields of Lake of the Woods, Manitou and Dryden," *Ont. Bur. Mines*, (1911) pp. 158–98, 26 figs., 4 pls.

[3] R. W. Seelye, "The Helen Mine, Michipicoten, Ontario," *Jour. Can. Min. Inst.*, XIII (1910), 121–34.

The ore body contains pockets of pyritic sand which have become more numerous with depth.

Wilson[1] reports on a reconnaissance survey of the area to the north of Lac Seul and east of Trout Lake in the Northwest Territories of Canada. He found only pre-Cambrian rocks, of which the oldest are chiefly hornblende schists and amphibolites. The greater portion of the area is underlaid by granites and gneisses intrusive into the basic rocks.

Wilson[2] finds that the pre-Cambrian rocks of the Algoma and Thunder Bay districts between 48° 30′ and 51° north latitude, and 84° and 87° 30′ west longitude, include Laurentian granite and gneiss with some hornblende schists, biotite schist, and diabase which may be Keewatin.

Wilson[3] states that the pre-Cambrian succession in the Nipigon river basin is as follows:

Keweenawan
- Diabase intrusives
- Dolomitic limestone
- Shale
- Sandstone
- Unconformity

Lower Huronian
- Small bands of conglomeratic arkose and slate
- Unconformity

Laurentian
- Granites and gneisses intrusive into Keewatin

Keewatin
- Greenstone and greenstone schists infolded with jaspilitic iron formation

Wilson[4] reports the following pre-Cambrian succession from the region about Larder Lake, Ontario and eastward:

Post Huronian
- Diabase, gabbro, porphyry, and lamprophyre
- Igneous contact

[1] A. W. G. Wilson, "Lac Seul to Cat Lake," *Canada Geol. Survey*, 1909, 23 pp.

[2] M. E. Wilson, "Geological Reconnaissance of a Portion of Algoma and Thunder Bay Districts, Ontario," *Canada Geol. Surv.*, 1909, 49 pls., 6 pls.

[3] A. W. G. Wilson, "Geology of the Nipigon Basin, Ont.," *Canada Geol. Surv. Mem. No. 1*, 1910, 152 pp., 16 pls., 4 figs., 1 map.

[4] M. E. Wilson, "Larder Lake and Eastward," *Canada Geol. Surv., Summary Rept.*, 1909, pp. 173–79, 1910.

Huronian
- Conglomerate
- Arkose
- Graywacke
- Conglomerate
- Unconformity

Laurentian
- Granite, gneiss, pegmatite, aplite
- Igneous contact

Keewatin
- Quartz-porphyry and porphyrite
- Rusty weathering carbonate rock
- Phyllite, slate, and graywacke
- Greenstone and green schist

Reviews

Climate and Evolution. By W. D. Matthew. Annals of the New York Academy of Science, Vol. XXIV, pp. 171–318.

This important paper is notable for the emphasis it lays on climatic variations and physical changes as agencies dominating organic evolution, for its adherence to the essential permanency of the continents, and for its unhesitating rejection of oceanic eversions and of extravagant bridge-building across abysmal depths for mere convenience in explaining biological distribution. In all these the author is loyal to the agencies attested by the geologic record and declines to go beyond them in summoning agencies of which the record gives no substantial authentication. He appeals to the powerful influence of climatic oscillations running back over the whole history of vertebrate life and beyond, whose verity is being constantly supported afresh by new evidence, and to the co-operative influence of physical changes connected with periodic diastrophism and denudation which have constantly varied the environment of life.

In addition to the long-recognized evidences of the essential permanency of the continents, the author cites the new support that springs from the geodetic evidences of isostasy and accepts all the difficulties this may seem to impose on the elucidation of biologic distribution. The author turns some of these supposed difficulties into evidences of the permanence of the ocean basins, by citing the obvious fact that a bridge between two great land masses should give a normal intermingling of the two faunas and floras inhabiting them and not a meager selection of forms susceptible of abnormal distribution by occasional modes of transportation. In this the author makes a notable contribution to the more critical study of what abnormal distribution really means. To the reviewer it seems probable that a really critical analysis of most cases of animal and plant distribution that are not in obvious harmony with the existing embossments and ridges, submerged or unsubmerged, will be found to imply selective transportation, not the normal commingling of species that naturally arises from a physical connection available to all species. The author's contribution to a closer scrutiny of the biological evidences of anomalous distribution is one of the first order of value. For its details the reader must consult the paper itself.

In his rejection of hypothetical eversions of ocean basins, the author does not hesitate to include Gondwana-Land, in some sense the parent—the mythical Atlantis aside—of the whole brood of lost continents. In this the author is likely to be sustained by stratigraphic and glacial evidence now already grown to a considerable mass which testifies, as well as bordering evidence well can, to the absence of any such elevated land connection between Australia, India, and South Africa across the Indian Ocean as is so commonly postulated to account for biologic distribution and the remarkable glaciation of those regions in Permo-Carboniferous times. In the opinion of the reviewer the stratigraphic and glacial evidences are not only quite against such intervening continent, but the hypothesis increases rather than alleviates the difficulties of explaining Permo-Carboniferous glaciation in those strange latitudes, and surely the difficulties are formidable enough without such hypothetical embarrassment.

The author does not push his negative attitude toward diastrophic extravagances farther back than Gondwana-Land because his study of vertebrate evolution does not seem to require it, but he might well have carried it back to the strange apparition of fishes and fishlike forms. Because coarse Devonian sediments with ichthyic faunas and terrestrial floras are found to encircle, in an imperfect way, the North Atlantic basin, an Atlantis is postulated in disregard of the strictly logical interpretation of both the physical and biological evidence which implies a disruption of the basin border, with a probable accentuation of the basin itself, instead of its eversion with the extremely improbable reversal of diastrophic action involved.

It is of course not improbable that some geomorphic changes of considerable importance have affected the ocean basins and the continental embossments in the course of geologic history, and possibly some of these may have been vital factors in the distribution of life, but the conservative example of the author in endeavoring to exhaust the probable influences of known fluctuations of milder type, attested by their own credentials, before resorting to collossal changes devoid of an appropriate record of their own, is wholesome and highly commendable. The paper should be read for its method as well as its material.

T. C. C.

The Noatak-Kobuk Region, Alaska. By PHILIP S. SMITH. Bull. U.S. Geol. Surv. No. 536, 1913. Pp. 157.

This report deals with the geology along the Noatak and Kobuk rivers which drain an area included approximately between 154° and 156° west longi-

tude, the Arctic Circle and 68° north latitude. The sedimentary rocks are divided into undifferentiated Paleozoic metamorphic schists, undifferentiated Paleozoic limestones, Silurian limestone, Upper Devonian limestone, Mississippian beds (Noatak sandstone, Lisburne limestone), a probable Jurassic Anaktuvuk group, a Lower Cretaceous or Upper Jurassic Koyukuk group, an Upper Cretaceous Bergman group, Eocene conglomerate, Pleistocene, and Recent. The igneous rocks consist of greenstones (pre-Carboniferous), probable Jurassic basic intrusives, flows, and tuffs, granitic intrusives (pre-Upper Cretaceous), and late Tertiary to Recent basaltic flows and volcanic ash. Deformation followed by erosion occurred previous to the deposition of the undifferentiated Paleozoic limestones. A period of deformation closed the Mississippian. The present topographic features were produced after the deposition of part of the Eocene. A series of east-west folds, broken by faults, affect the eastern part of the area, while north-south folds and faults are present in the western part. Auriferous gravels of notable production occur in the Shungnak and Squirrel River regions. There are gold and copper sulphide lodes, which have not as yet been developed to the producing stage.

V. O. T.

A Geological Reconnaissance of the Fairbanks Quadrangle, Alaska. By L. M. Prindle. With a Detailed Description of the Fairbanks District, by L. M. Prindle and F. J. Katz. And an Account of Lode Mining near Fairbanks, by Philip S. Smith. Bull. U.S. Geol. Surv. No. 525, 1913. Pp. 216.

The Fairbanks quadrangle, which forms a portion of the central plateau province of Alaska, lies between 64° and 66° north latitude and 146° and 150° west longitude. The most prominent structures are northeast-southwest trending close folds. The oldest rocks belong to the pre-Ordovician (?) Birch Creek schist; separated by an unconformity, a series of Paleozoic rocks (Ordovician to Carboniferous) follow. Unconformities also exist between the Silurian and early Devonian, between later Devonian and probable Pennsylvanian rocks, between the Upper Cretaceous and older rocks, and between the Upper Cretaceous and probable Eocene lignite-bearing beds. The occurrence of greenstones (basic volcanic lavas and tuffs), which are probably in the main of Devonian age, is interesting inasmuch as similar rocks appear in many parts of Alaska which are apparently of about the same age. The intrusion of granites and diorites succeeded the deposition of the Pennsylvanian strata. In one locality a basalt flow of probable Tertiary age was noted.

In the detailed description of the Fairbanks district are brought out the facts that this district lies wholly within the area of the Birch Creek schist, that great intrusions of diorite, granite, porphyries, and dikes occurred probably at the close of the Mesozoic, and that the auriferous and other deposits are genetically related to the intrusions. Gold placers have been developed on a

large scale in the Fairbanks district. These produced in 1912 gold to the value of $4,370,000 and silver to the value of $31,203. The development of the placers, according to F. J. Katz, has been chiefly by drift-mining methods, although open-cut mining is practiced in shallow ground. P. S. Smith in his discussion of lode mining notes that practically all of the developed veins are free-milling gold lodes, and that six properties are producing and have their own mills. The estimated production for 1912 from the lodes was $200,000. It is believed that this production will increase with the introduction of improved methods and more extensive development.

V. O. T.

The Koyukuk-Chandalar Region, Alaska. By A. G. MADDREN. Bull. U.S. Geol. Surv. No. 532, 1913. Pp. 116.

The Koyukuk-Chandalar region is bounded by 146° and 154° west longitude and the Arctic Circle and 68° north latitude. The main purpose of this report is to describe that portion of the area in which gold placers have been developed. The oldest strata are of sedimentary origin and Paleozoic age. These include the pre-Ordovician (?) Birch Creek schist, Devonian (?), and Carboniferous (?) beds. Cretaceous beds represent the Mesozoic sediments, and Tertiary and Quaternary deposits the Cenozoic. Granitic and dioritic intrusions (mostly of Mesozoic age) are associated with the Birch Creek schist; basic igneous rocks and tuffs occur with the Devonian (?); some volcanic rocks are present in the Cretaceous; the youngest igneous rocks are effusives, basaltic and andesitic lavas, and tuffs of Quaternary or Tertiary age. Unconformities are recognized between the pre-Ordovician (?) and Devonian (?); between the Carboniferous (?) and Cretaceous; between the Cretaceous and Tertiary; between the Tertiary sediments and the Quaternary or Tertiary igneous rocks; between these igneous rocks and the Pleistocene; and between the Pleistocene and Recent. The dominant structural lines trend a little north of east. The older schists are closely folded, while the Paleozoic series is thrown up into more open folds with many faults. The Mesozoic and Tertiary sediments are locally folded and faulted. Placer gold is, at present, the only mineral of commercial importance in the Koyukuk district. The gold occurs in the present stream deposits and bench deposits and has been derived chiefly from the Birch Creek schist. Surface mining methods are employed. In general, the yearly production of the Koyukuk district has gradually increased during the last ten years; the total estimated production up to 1912, inclusive, is $2,700,000. No lodes of commercial value have been found. In the Chandalar district probably "the most promise lies in its known quartz-lode gold deposits," since the placers are poor and local. The lodes are associated with diorite intrusions into the schist. Because of the difficulty of transportation of machinery and supplies, no producing mines are in operation.

V. O. T.

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SOME EVENTS IN THE EOGENE HISTORY OF THE PRESENT COASTAL AREA OF THE GULF OF MEXICO IN TEXAS AND MEXICO

E. T. DUMBLE
Houston, Texas

CONTENTS

EOGENE SECTION

The most complete section of the Eogene[1] deposits of the eastern border of North America is found in Alabama and Mississippi, but even this is somewhat indefinite in its upper members

[1] Used as inclusive of both Eocene and Oligocene.

and only by supplementing these with their marine representatives of the Florida region can the entire section be clearly understood

Thus constructed, the section would be:

	Alabama Mississippi	Florida
Oligocene:	Grand Gulf	Tampa
	Ocala	Chattahoochee
	Vicksburg	
Eocene:	Jackson	
	Claiborne	
	Wilcox	
	Midway	

The deposits as a whole are those of marine waters alternating with estuarine and fresh-water conditions, and the fossils indicate uniformly a warm or subtropical temperature of water.[1]

The time relations of the various members of these groups can be best made out in such an area of apparently continuous sedimentation as that found in this typical area, but when, as in this case, a study of the faunas of the various members indicate great physical changes intervening between them, further evidence of such changes is to be looked for in the extensions of the deposits. Some important evidence of this character is to be found in the Texas-Mexican region (Fig. 1).

CRETACEOUS-TERTIARY BARRIER

Prior to the deposition of the earliest Tertiary sediments in this area the Cretaceous rocks had been subjected to vulcanism, folding, and erosion. These movements began during the period of the Austin Chalk, at which time a barrier of some sort was created in the vicinity of the Salado and Sabinas rivers in northeastern Mexico which formed the beginning of what has been called the Rio Grande embayment. To the north of this barrier the succeeding Cretaceous sediments are largely clays and sands and include the coal beds of the Rio Grande region, while to the south we find only a great thickness of blue and black shales. The beds to the north are frequently very fossiliferous, while the shale beds are, so far as known, practically destitute of any trace of fossils.

[1] Dall, *Trans. Wag. Inst.*, p. 1540.

FIG. 1. Barrier in Gulf Coast region separating Atlantic and Pacific waters in Cretaceous-Tertiary times.

This movement was marked in the Texas area by Pilot Knob and other volcanoes south of Austin, which were active during the close of the Chalk deposition and the beginning of the Taylor marls, as is shown by the ash from them, which is interstratified with, and included in, these deposits.

At the close of the Cretaceous the movement was intensified and resulted in a land barrier which is now marked by the disconnected ranges and groups of hills that form the eastern border of the valley lying at the foot of the Mexican Cordilleras.

These groups and ranges include the San Antonio, San Juan, Vallecillo, Picachos, Papagallos, San Carlos, and Tamaulipas, for which, as a whole, Professor Cummins has proposed the name of the Tamaulipas range. Since the trend of the coast in this region is a little west of south, the southeast course of this Tamaulipas Range brings it rapidly nearer the Gulf. At Tordo Bay, some fifty miles north of Tampico, the main body of the hills is within ten miles of the coast, while outliers extend almost to the water's edge.

This range is made up of shales and limestones, more or less altered and disturbed by folding and by igneous rocks. No fossils have been found in the shales, but the limestones yielded a few *inocerami* and fragments of *Ammonites*. Dr. Stanton, who examined these, states that these limestones are not younger than the Taylor marls and may not be younger than the Austin Chalk.

The Tamaulipas Range represents the extension southward and culmination of the movement creating the Sabinas barrier, and just as the Sabinas barrier forms the southern limit and border of the Gulf Coast Cretaceous deposits, so does the Tamaulipas Range form the southern border and limit of the Gulf Coast Eocene. The difference in the character of deposits laid down in late Cretaceous time to the north and south of the Sabinas barrier is just as strongly shown in the difference between the Eocene deposits east of the Tamaulipas Range and those of the same age which are found to the west and south of it, since not only are the deposits themselves of different character, but even the fossils which the latter carry are entirely unlike those to the north.

CRETACEOUS-TERTIARY CONTACT

South of the Salado River the exposures of the Cretaceous shales along the line of contact with the Tertiary indicate that they have been subjected to more or less folding and to erosion prior to the incursion of the waters of the Eocene, so that the contacts from Rodriguez south show very decided unconformities.

On the Rio Grande the unconformity is shown partly by difference in direction of dip and partly by the fact that the Midway is found resting upon different members of the Escondido.

Northward from the Rio Grande the uppermost beds of the Cretaceous present, as a whole, a gentle slope toward the Gulf, and when we find contacts between these beds and the basal Tertiary the discordance of stratification is usually very small, even if it be present at all. There is, nevertheless, ample proof of folding in this area also, for logs of wells drilled in the Tertiary belt of western Louisiana and eastern Texas give positive evidence of broad folds in the underlying Cretaceous beds far to the seaward of their present outcrop.

The most striking feature of the orogenic movement in this area is, however, found in the numerous submerged hills of the salt domes that now occur as inliers in the Tertiary belt. These, as shown by the discordance of dip between the Cretaceous rocks and those of the surrounding Eocene, are pre-Tertiary.

LOWER EOCENE

Midway.—The initial sediments of the Eocene were marine and are fairly uniform in character throughout their entire extent within the area of the western Gulf Coast. While east of the Brazos they occupy a belt which has at times a width of several miles, with a total thickness of between 300 and 400 feet, west of that stream they are found in disconnected exposures due generally to the overlapping of later deposits.

This is clearly shown on the Rio Grande, where excellent contacts have been found by Kennedy, and where this substage attains a thickness of 300 feet. Here the Midway is frequently covered both by the Wilcox or Lignitic and by the Carrizo sands or basal Claiborne.

These conditions continue to the Salado River, but south of that stream the Midway covers a much larger surficial area and the Wilcox and Carrizo are less prominent.

By far the best exposures of the Midway which we have found are those in the valley of the Salinas River east of Ramones, Mexico. Here they occupy a considerable area and also form the Alto Colorado, a hill 250 feet in height which is composed entirely of Midway sediments. These have a total thickness of more than 400 feet. Among the fossils found in some abundance are *Ostrea pulaskensis*, *O. crenulimargo*, *Venericardia alticosta*, *V. planicosta*.

The last exposure of these beds within this area was found on the road to China, six miles southeast of Ramones.

That the close of the Midway was brought about or accompanied by an uplift is proven in the Rio Grande region, where it is seen to have a greater dip than the succeeding Lignitic series.

Wilcox or Lignitic series.—The series of fresh-water or estuarine deposits which followed the Midway shows little change in the general character of its sediments southwestward, with the exception that south of the Colorado it loses the larger part of the lignitic beds which are such a prominent feature of it north and east of that river.

North and east of the Colorado the Wilcox covers a broad area and has a thickness, as shown by wells drilled through it, of 1,000–1,200 feet. South of this river the exposure gradually becomes narrower and the beds thinner.

Its exposures along the Rio Grande are few and the thickness of the beds exposed probably does not exceed 100 feet. All of its observed contacts with the Midway were unconformable and there is a distinct difference in the degree of dip of the beds of the two stages. The sand and clays of the Wilcox Lignitic show evidence of emergence and erosion and in many places are entirely absent, having evidently been eroded prior to the deposition of the lowest Claiborne sands.

The Lignitic is found within a few miles south of the Rio Grande in a number of exposures. Farther south and beyond the Salinas it overlaps and covers the Midway, so that the latter

is seen no more and the Lignitic is itself in contact with the shales of the Cretaceous.

The most southerly exposure of the Lignitic was seen on the Conchos River a mile west of Angeles, where it is uncovered for a short distance by the erosion of that stream.

At the close of the Lignitic its sediments emerged and were strongly eroded prior to the incursion of the Claiborne sea.

In Lower Eocene time, therefore, the area north of the Tamaulipas Range was submerged and marine waters prevailed as far south as the latitude of Monterey. At the close of this period the marine waters receded and after a period of erosion were followed by the submergence of the deposits under fresh or brackish waters. The close of this second or brackish-water period of sedimentation was followed by its emergence and erosion.

In the region west and south of the Tamaulipas Range no deposits have been found which can certainly be referred to the Lower Eocene as represented north of that range by the Midway and Wilcox.

Beginning near the foot of the high plains of the interior of Mexico there is a series of blue and gray shales which have a wide development, forming a large part of the surface rock of the valley between the Cordilleras and the Tamaulipas Range and continuing southward beyond the Tuxpam River as far as examinations were carried. In the northern part of this area these shales have been traced to a direct connection with those found in the San Juan and Papagallos mountains, which, by their relations to the basal Tertiary and the fossils found in limestones immediately underlying them, were proved to be Upper Cretaceous.

There is undoubtedly a great thickness of these shales, but no fossils have ever been found in them. There is, therefore, a possibility that in their upper portion some representative of early Eocene time may be present, but, if so, it has not yet been differentiated.

So far as yet recorded no fossils characteristic of the Midway or Lignitic have been found between Tordo Bay and Yucatan, although fossils of both periods are known to exist in Venezuela and elsewhere to the south.

MIDDLE EOCENE

Buhrstone and Claiborne.—In the Alabama section the Buhrstone and Claiborne have a total thickness of about 450 feet, the Claiborne proper being only one-third of this amount. The surface exposures rarely exceed a total width of fifteen miles. In Texas the narrowest portion of this Claiborne belt occurs on the Colorado River and has a width of twenty-five miles, which widens northeast to ninety miles on the Neches and attains a similar width southwestward in the Rio Grande region. Along the Rio Grande, which crosses the formation obliquely to the dip, the belt has an exposure of 150 miles. The beds in Texas also have a correspondingly increased thickness and have been divided as follows: the Carrizo sands corresponding to the Buhrstone; the Marine, Yegua, Fayette, and Frio substages—all of which carry fossils characteristic of the Lower Claiborne. These substages are present from the Conchos or Presas River in Mexico to the Colorado in Texas. The Frio has not been recognized east of the Colorado, and east of the Neches the Fayette is also wanting, except as isolated patches, owing to pre-Jackson erosion.

The Carrizo sands which mark the beginning of the Middle Eocene deposition are the stratigraphic continuation of the Queen City beds of northeast Texas. Throughout this region where Kennedy first observed them, the relations of these sands to the underlying Lignitic is such that he considered them to be the upper member of that stage and so described them. Later examinations in the western part of the state, however, brought out the fact that the Carrizo sands were entirely unconformable with the Lignitic deposits and that in some localities they grade upward into the basal beds of the Marine substage.

In Northeastern Texas the Carrizo sands have a thickness of 70 feet and appear to be conformable with the Lignitic. On the Atascosa the contact was not observed, but the sands were apparently of much greater thickness. On the Rio Grande the beds are exposed over a very wide area, outcropping along the river for a distance of forty miles. They have a less marked dip than either the Lignitic or Midway and are found resting on both of these and

overlapping them to contacts with the Escondido beds and the Papagallos shales of the Cretaceous.

South of the Rio Grande the Carrizo sands are well exposed and form the crest and eastern slope of various small groups and ranges of hills lying between the Rio Grande and Azulejo and of the larger range south of that point, known as the Ceja Madre. These hills have a total length of about ninety miles and reach nearly to the Salado River.

South of the Salado the Carrizo is not so prominent and was observed only in a few localities.

The Marine beds in places, as on the Rio Grande and the Atascosa, are directly connected with the Carrizo sands by transitional beds of lignitic shales, but at other places, and especially in east Texas, no such connection can be found and beds of greensand or of brown sandstone are directly superposed upon the Carrizo, although usually without apparent unconformity. Local movements during Marine time are, however, registered in the sediments in some parts of the area.

The Marine beds have a thickness of 800–1,000 feet. The eastern portion of the deposits seems to have been most highly glauconitic, although the beds everywhere give evidence of the presence of this mineral in greater or less quantity.

In the eastern portion of the state the alteration of this glauconite has resulted in the formation of extensive deposits of excellent iron ores, but in the western portion the resulting ferruginous matter is widely disseminated and simply forms a cement for the brown ferruginous sandstone which covers such a great area in Southwestern Texas.

On the Rio Grande the carbonaceous shales of the Carrizo find their culmination in the San Pedro and San Tomas beds of cannel coal between which are beds bearing characteristic Claiborne fossils.

The exposure of the Marine beds on the Rio Grande has a width of more than fifty miles. They narrow rapidly to the south and where they cross the Salado west of Guerrero the outcrop is not more than five or six miles wide.

They are well shown in the valley of the San Juan and are traceable for twenty-five miles south of that stream, when they are covered by the Yegua. They are not observed on the Conchos.

Fossils are abundant throughout the entire extent of the substage, although the upper portion is usually the most fossiliferous portion. In the Texas area in addition to the characteristic fauna we find a few forms practically identical with those of the Tejon. In the San Juan region, associated with the *Venericardia planicosta*, there are found vast numbers of a varietal form very similar to, if not identical with, *V. potopacoenses* of the Maryland Eocene.

The succeeding substage, the Yegua clays, appears to be unconformable with the Marine. The Yegua deposits are predominatingly lignitic clays, but comprise beds of sands and greensands as well. They, unlike the earlier Lignitic, maintain their thickness of 1,000–1,400 feet and carry their deposits of lignite entirely across the state. They are also the principal gas-bearing beds of the Eocene. Like the lower Lignitic, they become more sandy on the Rio Grande and this condition persists throughout exposures south of the river to Mendez on the Conchos or Presas.

In their eastern part these beds are not as fossiliferous as they are west of the Guadalupe, although they carry fossils in many places, especially between the Brazos and the Colorado. On the Rio Grande they are nearly as fossiliferous as the Marine.

It has been suggested that the so-called Cocksfield Ferry beds on the Sabine might be of Upper Claiborne age, but they have been connected by careful tracing with the basal Yegua and are, therefore, Lower Claiborne.

To these deposits of Yegua clay there succeeded another substage, the Fayette, which, while it carries beds of lignitic clays and sands with some lignite, is principally composed of sands and sandstones. Some of the sands are friable while others are quartzitic and they carry a considerable amount of fossil wood which is opalized instead of being simply silicified. They have a thickness of 500–600 feet.

East of the Guadalupe, with rare exception, the Fayette does not carry any fossils except the wood, but west of that stream it contains the large oyster, *O. alabamiensis var. contracta* of Conrad,

which is one of its most characteristic forms. On the Nueces and Rio Grande we find, besides this oyster, numerous other forms which link it positively with the Claiborne. This is especially true of the exposures on the Rio Grande which have a width of fully forty miles with many fossil beds. These include:

Venericardia planicosta, Lam.
Anomia ephippioides, Gabb.
Leda opulenta, Con.
Crassatella protexta, Con.
Tellina mooreana, Gabb.
Cytherea bastropensis, Har.
Conus sauridens, Con.
Volutilithes petrosus, Con.
Levifusus trabeatoides, Har.
Lacinia alveata, Con.
Pseudoliva vetusta, Con.
Turritella nasuta, Gabb.

South of the Rio Grande it is found in a broad belt extending southeastward by way of Mier, Aldamos, China, and Mendez to the Conchos and beyond as far as the Chorreras, where it finally passes under cover of the Oligocene.

Between the Conchos and the Chorreras the Fayette apparently forms a portion of Mt. Corcovada east of Burgos, being elevated by a basaltic extrusion.

The Frio clays which follow are found overlying the Fayette sands from the Conchos River to the Guadalupe. On the Conchos and on the Ramones Mountains lying north of it they show a considerable thickness of yellow and darker clays which weather white and are accompanied by considerable quantities of disseminated gypsum and also beds of this material. Farther north the Frio beds are not so thick and in many places only a few feet of the formation are found between the sands of the Fayette and the overlying Oakville sands of Miocene age. Such fossils as have been found are apparently of Claiborne age and the clays are, therefore, referred to that stage.

The Middle Eocene of this Gulf area represents a succession of alternating marine, lagunal, and swamp conditions. The marine conditions were most constant in the Rio Grande region and farther

south, while in the eastern portion of the state the lagunal and swamp conditions apparently predominated.

There are unconformities here and there and evidences of local warping of the strata in the various substages, but they are, nevertheless, bound together by transitional beds or identity of fossils.

The constantly changing conditions of deposition in the region east and north of the Tamaulipas range during Middle Eocene time, as shown by the variableness of the deposits above described, did not extend to the region lying west and south of that range.

West of the Tamaulipas Range the blue shale of the uppermost Cretaceous is followed by other shales so similar in general character that only the finding of Eocene fossils in them suggested the differentiation of the two.

The only localities where exposures of these beds have certainly been found is in the vicinity of Alazan, on the Buena Vista River, about twenty-five miles northwest of Tuxpam.

The deposits are bluish shales that weather white or yellow and their exact relations to the Cretaceous were not determined, on account of the inadequate exposures. The fossils are widely scattered through the shale and it is possible to collect them in good condition only from decomposed fragments of the rock.

This locality was discovered by Mr. DeGolyer and fossils found by him were recognized by Dr. Dall as Pacific Coast forms. It was afterward visited by Professor Cummins, who made a somewhat larger collection of the fossils. These were also submitted to Dr. Dall and afterward turned over to Professor Roy E. Dickerson for study in connection with other Pacific Coast faunas of Eocene age in his work at the University of California.

The collection of fossils made at this locality by Professor Cummins comprises a fauna consisting principally of gasteropods and small bivalves. The preliminary examination failed to show any forms which could be referred to our Claiborne, but did show some that were apparently identical with, and many that were very similar to, Eocene Tejon of California.

These beds dip eastwardly and probably extend some distance north and south of Alazan. Shales similar to these were observed along the railroad north of the Panuco River and on the San

Antonio River west of Cruz, a station on the railway between Tampico and Monterey. Cummins correlates these shales north of Panuco and west of Cruz with the Alazan deposits because of common lithological character, although he found no fossils in them.

They were, however, certainly found in drilling a well at Topila, on the Panuco River, at a depth of 1,810 feet, as shown by the fossils brought up with the drill. Of the fifteen species collected at this locality nearly one-half are identified as Tejon species and the others as nearly related to them.[1]

UPPER EOCENE

So far we have not positively recognized the Frio clays, which are the uppermost beds referable to the Middle Eocene, east of the Colorado. Other beds of clay have been referred to this substage, but closer investigation seems to show that the Frio clays are not present in eastern Texas. Growing out of such correlation it has been presumed that both the Fayette and Frio of eastern Texas are later than the Lower Claiborne and correspond with the Jackson and the Oligocene.

In the area west of the Colorado River, where these beds have had closest study and where they are most fossiliferous, it is very certain that this presumption is erroneous. There is a continuity of the Lower Claiborne fauna through these upper beds and no forms whatever of Jackson or later age have been observed.

In all this area the Frio is found resting seemingly upon an uneroded surface of Fayette sands and north of the Rio Grande it is unconformably overlain by the Oakville Miocene without beds of either Upper Eocene or Oligocene age intervening.

East of the Guadalupe River the conditions are somewhat different. In the valley of the Colorado and eastward to the Sabine it is evident that subsequent to the deposition of the Frio and prior to the beginning of Jackson deposition as now known, there was a period of erosion during which the Frio, more or less of the Fayette, and even a part of the Yegua were carried away. This erosion probably took place during the deposition of the Upper

[1] *Science*, XXXV, 906–8.

Claiborne sediments in other regions and appears not to have greatly affected that portion of our Claiborne belt lying between the Guadalupe in Texas and the Conchos in Mexico. During these oscillations of the land it is evident that the relief at any time was slight.

Jackson.—Beginning west of Groveton, in Trinity County, Texas, in the valley of the Neches, and stretching eastward to the Sabine is a belt eight or ten miles in width in which are found calcareous clays and sands with nodules and bands of impure limestone, carrying fossils of Jackson age. For the most part these deposits lie directly upon the clays of the Yegua, since only occasionally are the Fayette sands to be found between them.

The identity of these deposits is clearly evident from the considerable number of fossils found at various localities.

The Jackson sediments have not been found west of Groveton. In their place, however, the Fayette sands gradually assume their normal position in the section, showing less and less erosion, until the Colorado drainage is reached.

The Jackson shore line may have extended westward as far as the Brazos, for sands thought to be at the top of the Fayette near Wellborn have yielded fossils which are now said to be of Jackson age, although originally the collections were classed as Lower Claiborne.

These are supposed to be, however, a lower horizon in the Jackson than the deposits farther east.

In southeastern Texas we have found indications of the gulfward extension of the Jackson deposits in fossils from oil wells drilled at Saratoga and Sour Lake, but southwest of the Brazos no indications whatever of any materials of this age are known either in Texas or in Mexico.

We have, therefore, in this eastern Texas area simply the western limit of the Jackson deposition and the materials themselves in many places give evidence of the near-shore conditions.

Almost nothing has been found in this area that throws light on the happenings between the close of the Middle and Upper Eocene and the beginning of Oligocene deposition. Whatever there may have been is covered by the succeeding beds of the Oakville.

OLIGOCENE

Nowhere within the Texas area has any trace of Vicksburg strata been found, but Hilgard and Loughridge both contended that the Grand Gulf sandstones were found along the Sabine, and Loughridge traced them nearly across the state.

There is a belt of sandstone crossing the state just as described by this observer and it has some of the lithologic characteristics of the Grand Gulf. Our investigations seem to indicate, however, that while it may be traced in apparent continuity across the area it is in reality not homogenous, but is in fact a composite series. That it represents in part remnants of a deposit of the Grand Gulf sands, even if not fully susceptible of proof, is strongly probable. We also know that it comprises beds of later deposition which have surrounded and overlapped these Grand Gulf remnants. These later deposits are the Oakville sands of western Texas and are of Miocene age. Among the outcrops now remaining which we think are Grand Gulf are those at Colmesneil, the quarry north of Corrigan, near Riverside, etc.

The lack of fossils in the Grand Gulf sands and the close resemblance to them (even including the quartzitic phases) of those of Oakville age will probably prevent any complete differentiation of the two, although this may be done in part through a study of the scanty flora occurring in them here and there.

If any other Oligocene deposits besides the remnants of the Grand Gulf were laid down in the Texas area they have been eroded or are also covered by the overlap of the Oakville.

In Mexico, on the contrary, we find a broad development of the Marine facies of the Oligocene deposits such as are found in Florida and the Antilles as distinguished from the near shore brackish-water phase of most of the Gulf area.

North of the Tamaulipas Range we find, overlying the Eocene, a series of yellow sands, clays, and calcareous beds which carry an abundant Oligocene fauna. They were first studied in the vicinity of San Fernando, on the Conchos River, where there are fine exposures, and we have called this local development of the beds the San Fernando.

These Oligocene deposits are found resting upon the Frio clays along the Conchos River and the eastern front of the Pomeranes Range, and they extend northward to within fifty miles of the Rio Grande, beyond which point we find no trace of them, possibly because of the extensive development of the Reynosa surficial deposits in that region.

South of Abasola the Oligocene overlaps the Frio and is in direct contact with the Papagallos shales on the west and extends eastward to the Gulf shore.

The corals found in some of these beds have been studied by Dr. Vaughan, who says that they indicate an Upper Oligocene horizon, about equivalent to the Chattahoochee of Georgia. This is borne out by such of the molluscan remains as have been examined, and so far our collections from this area have not yielded any forms characteristic of the Lower Oligocene. Among the corals determined are: *Favites* (?) *polygonales* Duncan, *Goniastrea antiquaensis* Duncan, *Acropara* ? sp., *Orbicella cellulosa* Duncan, *Orbicella* n. sp., *Geneopora* sp. very similar to, or identical with, an Antiguan species.

South of the Tamaulipas Range the yellow clays and sands of the Oligocene cover a broad area. Between the Buena Vista River and the Tamiahua Lagoon many exposures are found and the beds are highly fossiliferous. *Cristalleria* and *Nummulites* abound and *Orbitoides papyracea* Bou. is found from the Buena Vista to the Tancochin at Cerro del Oro. Dr. Dall states that this form occurs in the Lower Oligocene beds in Alabama, indicating a Lower Oligocene age for this portion of the deposits. This Oligocene belt has a known width of twenty miles east of the Buena Vista River and has been observed twenty miles west of that stream resting upon the shales of the Cretaceous.

In addition to the fossils named there are many others, including several echinoderms, pectens, and other bivalves and numerous gasteropods.

While a detailed examination of these beds has not been made, enough has been done to prove that there is in this part of Mexico very considerable thickness of these Oligocene deposits and that they represent portions, if not all, of both Lower and Upper Oli-

gocene time. The lower beds, so far as now determined, are confined to the region south of the Panuco, while the upper stretch northward beyond the old Eocene barrier nearly to the Rio Grande.

MIOCENE

In Texas, as we have seen, such Oligocene deposits as may occur are overlain directly by the Miocene (Oakville) and the sands of one are almost indistinguishable from the sands of the other, both being probably of fresh- or brackish-water deposition.

In Mexico the yellow clays and sands of the Oligocene are also overlain by similar yellow clays and sands of Miocene age and the deposits themselves are of very similar character and thus far no evidence of a decided break in sedimentation between them has been found, the only change being in the fossils contained in them, those of the Miocene being characteristically developed in the beds around Tuxpam.

TAMPICO EMBAYMENT

The importance of the Tamaulipas Range as a structural feature is, therefore, clearly shown, and it is now apparent that at the beginning of the Tertiary and through the Eocene the waters of the Mexican Gulf covered only that portion of its present coast line in Mexico which lies north of Tordo Bay about 225 miles south of the mouth of the Rio Grande.

It would also appear that the waters of the Gulf and those of the embayment south of the Tamaulipas Range, although they were evidently separated by a comparatively narrow strip of land, did not have such a connection as would permit a commingling of the two faunas, since they have only four or five forms in common.

The Topila well mentioned above is less than twenty miles from the present shore line of the Gulf of Mexico and is much nearer to its present waters than is any bed of similar age in the region north of the Tamaulipas Range. There is no indication of a barrier to prevent the extension of these deeply buried Alazan beds with their Pacific fauna eastward or southeastward until they actually underlie the present waters of the Gulf. It is clearly evident, therefore, that during the Middle Eocene the Gulf waters did not reach below Tordo Bay on its present coast line, but were

held by a barrier or the southeastward extension of the Tamaulipas Range and that part of its present coast in the vicinity of Tampico was occupied by the waters of the Pacific.

The observed facts indicate that during the Middle Eocene the waters of the Atlantic occupied the territory north of the Tamaulipas Range and the waters of the Pacific the region south of that range, and that with the beginning of the Oligocene, the waters of the Atlantic claimed both of these regions. It is therefore evident that the connection between the Pacific Ocean and its Tampico embayment was closed entirely, either during the Upper Eocene or at its close, and that the extension of the present Gulf coast southward from Tordo Bay dates from this period.

While the exact area embraced in this embayment south of the Tamaulipas Range is unknown, its point of connection with the Pacific Ocean was probably near the present Isthmus of Tehuantepec and its western border was formed by the escarpment of the Cordilleras. It is probable that its northerly extension occupied the site of the valley now traversed by the railroad between Tampico and Monterey. The Tamaulipas Range must have formed its eastern boundary in this northern portion, but we have, at present, no idea of its eastern limit farther south beneath the western margin of the present Gulf of Mexico. It seems probable, however, that this Tampico embayment was a body of water not unlike the Gulf of California, both in extent and in trend, and that it made a peninsula of that part of the Republic of Mexico north of the Isthmus of Tehuantepec just as the Gulf of California now makes the peninsula of Lower California.

THE STRENGTH OF THE EARTH'S CRUST—*Concluded*

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PART VIII. PHYSICAL CONDITIONS CONTROLLING THE NATURE OF LITHOSPHERE AND ASTHENOSPHERE

SECTION B

RELATIONS WITH OTHER FIELDS OF GEOPHYSICS

RELATIONS WITH OTHER FIELDS OF GEOPHYSICS

Erroneous Conclusions Reached by the Rectilinear Projection of Surface Conditions

Geologists early became aware that temperature increased with depth. Projecting this gradient as a straight line indicated that at no great depth the temperature was sufficiently high to melt all rocks and, in testimony, volcanoes brought such melted rocks to the surface. The earth was consequently looked upon as a molten or even gaseous body enveloped by a thin crust of solid rock. The logic of this conclusion seemed incontrovertible and moreover it was in accord with the simpler expectations from the nebular hypothesis. Nevertheless, direct and positive evidence from several independent sources has forced on geologists the belief that the earth is not only solid throughout, but, as a whole, is more rigid than steel. Slowly and with difficulty the older view has therefore had to be abandoned. Yet it continually recurs in one form or another, advocated chiefly by writers who see the direction in which the surface evidence of temperature gradient leads, who regard it as compulsory, and who do not recognize or give equal

weight to the direct evidence regarding the nature of the earth's interior. Because of the ease and certainty of laboratory studies there is a tendency to treat the interior of the earth as though it were incapable of speaking for itself through the evidence of geophysics, geodesy, and geology, but must remain forever a playground for the speculative imagination. Largely unknown the nature of the earth's interior is and long must be; laboratory studies on the influences of heat, of pressure, and of chemical composition, upon the physical state of the crust, must constitute the paths which guide in the search downward into the unknown; but the final test of hypothesis must be the direct testimony of the earth itself.

The rectilinear projection of surface conditions is based on the assumption that the temperature gradient is a straight line to great depths, or that strength, or density, or porosity, as the case may be, is not changed by the pressures of the interior. Such assumptions lead to views more or less in opposition to those reached in the present investigation. They must, therefore, be discussed to some degree. An illustration of these dangers of reasoning by unchecked extrapolation is supplied by a paper written by Arrhenius,[1] selected for discussion because of the eminence of the author in the fields of physics and chemistry, the definiteness with which his conclusions are stated, and the wide citation which this paper has achieved.[2] In this paper the arguments are given in favor of a gaseous nature of the interior of the earth, carrying forward an idea first suggested by A. Ritter in a series of "Researches on the Height of the Atmosphere and the Constitution of Gaseous Heavenly Bodies."

From the rate of increase of temperature with depth Arrhenius argues that at a depth of 40 km. the crust must pass into a molten condition, but one which, because of pressure, is a viscous and highly incompressible liquid. At a depth of some 300 km. the temperature, he states, must be above the critical temperature of

[1] "Zur Physik des Vulkanismus," *Geol. Foren. i Stockholm, Forhandl.*, XXII (1900), 395–419.

[2] See for example its presentation by A. Geikie, *Text Book of Geology*, Vol. I (1903), pp. 71–74.

all known substances, and therefore the liquid magma passes into a gaseous magma extending to the center of the earth. The author then notes that the chemical elements of highest atomic weight are not detected in the sun, but states that without doubt they occur, and concludes from this that they must be concentrated by virtue of gravity toward the sun's center.[1] The high density of the earth's interior is accordingly to be explained by the presence of substances heavier than surface rocks. For many reasons, as the dominance of iron in nature, as shown by meteorites, by the spectrum of the sun, and by the magnetism of the earth, it is to be concluded that this substance which he thinks necessary to account for the high density of the earth's interior is metallic iron. The earth consists consequently of the following portions measured from the center on the radius. Eighty per cent of the radius is gaseous iron, 15 per cent is gaseous rock magma, about 4 per cent is fluid rock magma, and somewhat less than one per cent is solid crust.[2]

To reconcile these conclusions with the incontrovertible evidence of rigidity, Arrhenius takes up another line of rectilinear extrapolation and carries it to an equally extreme degree. Fluids in general show a somewhat readier compressibility than solids. At high pressures then it is argued that liquids will customarily occupy less volume than solids and the pressure will tend to lower, not raise, the melting-point. Consequently, the rigidity cannot be accounted for by the maintenance of solidity through pressure. The author then points out that under enormous pressures all substances, even gases, must become highly incompressible; and that at high temperatures, where the volume is maintained the same, the viscosity of gases or fluids increases with increase in temperature. From this it is argued that in the central parts of the earth gaseous iron is more incompressible and viscous than solid steel. It is by enormous pressure consequently in spite of a gaseous nature that the interior of the earth exhibits its great rigidity.

Vulcanism according to Arrhenius is connected with the free seepage of ocean water downward through the crust which, he holds, constitutes a semipermeable membrane. By the absorp-

[1] *Op. cit.*, p. 402. [2] *Op. cit.*, p. 405.

tion of water into the heated rocks the conditions for volcanic activity are initiated. This argument, like the others, is in the form of a great extension or extrapolation of factors operative in a small way in the laboratory to conditions in nature which are wholly different in magnitude.

As comments upon this paper, it should be noted that nearly every conclusion applying to the sun and earth may be questioned.

At a depth of 1,000 km., according to Arrhenius, the temperature is about 30,000° C. The gradient is thus taken as essentially a straight line from the surface downward. There is no demonstration as to why this rectilinear extension is assumed, whether it is to be regarded as an adiabatic temperature curve produced by condensation under pressure or produced in some other way. The influence of cooling through geologic time in changing the outer gradient is not considered; nor the influence of rising magmas. The existence of radioactivity was then just beginning to be appreciated and naturally could not have been evaluated, but the data for a discussion of the other factors, though at hand, was neglected.

There is no demonstration that the heavy elements are concentrated in the sun's interior, or that the earth is mostly metallic iron. It is possible that the earth is thus constituted, but it must be proved on better evidence than a citation of the dominance of iron in nature. The incompressibility of all substances, both fluid and liquid, increases greatly with great increase of pressure, following apparently parabolic curves. Therefore, it cannot be argued with any assurance that the high incompressibility of the earth's interior proves the presence of iron, or that under such pressures the fluid occupies less volume than the solid state.

At a depth of 1,000 km. Arrhenius states that the temperature is about 30,000° C. and the pressure 250,000 atmospheres.[1] If, under these conditions of exalted temperatures, gaseous rock or iron has a viscosity equal to that of solid steel it may well be asked how the stars, with their immensely greater masses and consequent internal pressures, can maintain a convective circulation competent to keep up their enormous surface radiation. Furthermore, however viscous a compressed gas or liquid may be, this property

[1] *Op. cit.*, p. 400.

should be distinguished from rigidity. If a body can resist even small shearing stresses for an indefinite period, it has the essential properties of a solid and not a gas. If it possesses real rigidity, even if it should be true that under relief from pressure the substance would turn into a gas, yet such relief cannot take place and it is a confusion of terms to speak of the substance as a gas when exhibiting to a striking degree the essential qualities of a solid. This distinction between viscosity and rigidity is of first importance, yet is not mentioned by Arrhenius. Although undercooling of a fluid into a glass gives rise to the elastic properties of a solid, it has not been shown that increase of pressure, however great, upon a gas above the critical temperature would transform increasing fluid viscosity into solid rigidity and plasticity such as is exhibited by the earth.

As to the hypothesis that the crust is a semipermeable membrane, permitting a free downward seepage of ocean water, but little need be said, since this is a subject which has been much discussed in recent years and is now largely discarded by geologists. The evidence against it is varied. Petrologic study shows the deep rocks to be impermeable and unaltered; beyond a shallow depth they are dry, and their gaseous and liquid occlusions are held unchanged for geologic ages. Unsound conclusions have been built upon the behavior of steam within porous sandstones, combined with confusion of the rate of diffusion under enormous pressure-gradients in the laboratory with enormous pressures, but low pressure-gradients within the crust. Furthermore volcanoes are not restricted to the vicinity of the sea and their emanations are not of the proper composition to have been derived from ocean waters. As Suess has said, volcanoes are not nourished by the sea, but every volcanic eruption adds to the waters of the ocean.

The paper under discussion was written by a scientist who has done much exact work in physical chemistry, but who in passing to geologic thinking has adopted the habit of an earlier generation —a habit of speculative thought, suggested by chemical and physical concepts and not verified by a study of the earth. The form of present geologic investigations has, however, advanced to the quantitative stage, although the data are often so inexact that the

order of magnitude, or the direction of the truth, is all which may be now ascertainable.

In conclusion, it is seen that the hypotheses outlined by Arrhenius imply a thinness of the crystalline lithosphere and a crustal weakness wholly at variance with the conclusions regarding strength which have been reached in this investigation. They imply a difference in nature of the earth's interior from that given by the more direct lines of evidence, as shown by the body resistance of the earth to vibratory distortions of both short and long periods. Because of these many difficulties, this group of hypotheses, adopted by Arrhenius, has already been largely discarded, though they still find considerable acceptance, more especially by workers in related fields of science. But the measures of lithospheric depth and strength which appear to be given by geodesy add their testimony to the cumulative evidence against these views.

THE EVIDENCE OF TIDES ON RIGIDITY AND STRENGTH

The tidal distortion of the solid earth measured by means of the horizontal pendulum has shown that its rigidity is of the order of magnitude of steel. But the recent measurements by Michelson and others, employing a long horizontal pipe partly filled with water, showed clearly that the earth's rigidity is even greater than that of steel.[1] This higher value is in agreement with the inductions from the observations on the variations of latitude. But these measurements give the rigidity of the earth as a whole, not the distribution of rigidity. The resistance to tidal deformation is furthermore complicated by the influence of gravity and increasing density with greater depth. Even if the earth were a liquid globe it would resist tidal distortion to one-third the degree of the resistance of a globe of steel, and if the liquid sphere were denser inside, this ratio would be further decreased.[2] Notwithstanding this factor, however, it is clear that the earth as a whole is more rigid than steel. As the outer part is known to be less rigid than steel, it follows that the rigidity of much of the interior must be

[1] "Preliminary Results of Measurements of the Rigidity of the Earth," *Jour. Geol.*, XXII (1914), 118.

[2] A. E. H. Love, *Elasticity*, p. 306.

proportionately higher. But the tidal stresses, though serving as a measure of the rigidity of the earth as a whole, are so small that they are ineffective as a measure of the strength of the earth as a whole, or of even its weakest parts. The smallness of the stresses can be appreciated by noting Darwin's numerical calculations. In his original paper Darwin arrived at the conclusion that the tidal stress-differences at the center of the earth were eight times as great as at the surface, and this result has been widely quoted. In the final publication, however, a correction is made showing that this is the ratio between the surface stress *at the poles* as compared to the center. The stresses at the poles, at the equator, and at the center he finds to be in the ratio of 1 to 3 to 8. The diurnal tide gives an actual stress-difference per square centimeter amounting to 16 grams at the poles, 48 at the equator, and 128 at the earth's center.[1] The strength of granite at the surface of the earth averages about 1,700,000–2,000,000 grams per square centimeter. The elastic limit for steel subjected to tensile or compressive stresses in one direction ranges from about 3,500,000 grams to 4,500,000 grams per square centimeter, according to the grade of the metal. The ultimate strength is about twice as high as the elastic limit. Thus the earth is stressed by the tidal forces even at the center to only about one part in fifteen thousand of the strength of good granite at the surface, or about one part in twenty-seven to thirty-five thousand of the limits of perfect elasticity which steel exhibits in the laboratory. With stresses so small it is not surprising that although tides give measurements of rigidity their evidence regarding viscosity is most uncertain. The results of estimates of the viscosity are more or less contradictory and so small as to be within the probable error of determination. Nevertheless Schweydar considers that there is a suggestion of a slightly plastic zone extending from a depth of about 120 to 620 km. Although this has been adopted in the present article as the limit of the asthenosphere, it would appear that the convincing proof for the existence of such a zone, and the determination of its limits

[1] George H. Darwin, "On the Stresses Caused in the Interior of the Earth by the Weight of Continents and Mountains," *Collected Scientific Papers* (1908), II, p. 481; original publications, *Phil. Trans. Roy. Soc.*, CLXXIII (1882), 187–223, and *Proc. Roy. Soc.*, XXXVIII (1885), 322–28.

also is more likely to be given by the geologic and geodetic evidence rather than from that yielded by the tides, provided that the present hypothesis of the existence of an asthenosphere is accepted.

It might seem that if the asthenosphere is strained to its limit by permanent stress and is slowly yielding, that even the small and rhythmic tidal stresses, like the last straw on the camel's back, might reveal a lack of resilience in the region of yielding. The distinction was emphasized in Section A, however, that an elastic limit which is determined for permanent stress by a facility of recrystallization at a high temperature may be a far lower elastic limit than that which would exist for rapid rhythmic stresses. Recrystallization would theoretically go forward a little more rapidly during the additive phase of the tidal stress, but the process is presumably so slow, and the tidal stress so small and rapid, that no appreciable effects would be attained before the following of the negative phase. A high resilience of the earth under tidal stress seems therefore quite compatible with the existence of a slowly yielding asthenosphere.

THE EVIDENCE OF EARTHQUAKE WAVES ON RIGIDITY AND DENSITY

The speed of an elastic wave through a solid varies directly with the square root of the modulus of elasticity and inversely with the square root of the density. There are two waves, corresponding to the elasticities of volume and form respectively, the one measured by the modulus of compressibility, the other by the modulus of rigidity. The first is the longitudinal or radial wave, the second is the transverse wave. The former outruns the latter and gives rise to the first preliminary tremor by which the earthquake records itself in distant regions. The transverse vibration is felt as the second preliminary tremor, followed by the much larger oscillations of the principal wave. The first two go through the earth, the latter passes around the surface. The fact that there is a transverse wave shows that the earth is solid throughout. But the vibrations at the point of emergence for waves which have penetrated more than half-way into the earth are so faint because of distance that their beginnings are in doubt, and consequently the speeds of transmission below one-half of the radius are uncer-

tain. These greater depths do not, however, so immediately concern the present subject. For the outer quarter of the earth both radial and transverse waves increase in velocity of transmission with depth, showing that incompressibility and rigidity increase faster than density and reach values greater than those exhibited by steel at the surface of the earth.[1]

So much is certain, but when it comes to testing the character of any particular shell by means of the velocities and character of the vibrations which have passed through it, there is but little certainty. The difficulty of an exact interpretation is discussed well by Knott.[2] To illustrate the variety of opinions, Benndorf has worked out a law according to which the speed of transmission increases rapidly to a depth of 200 miles (320 km.) from the surface. Knott assumes a constancy of speed below a depth of 400 miles (644 km.).[3] Wiechert has concluded that there are sudden changes in velocity at depths of 1200, 1650, and 2450 km. Poisson's ratio which expresses the relationships of the elasticities of form and volume remains, however, practically constant throughout, having a mean value of 0.27.[4] These changes imply surfaces of discontinuity. If real, however, they are deeper than the shells of the earth involved in the problems of isostasy. The conclusions rest, however, upon data of doubtful reliability. Reid has made a critical examination of this subject in connection with his comprehensive study of the excellent records obtained from many parts of the world of the California earthquake of 1906.[5] Following Wiechert's method, the curves representing the normals to the wave fronts and the velocities at various depths were computed from the data of the seismograms. The result showed that for the radial or longitudinal wave the velocity increased rapidly with depth but with decreasing rapidity, from 7.2 km. per second at the

[1] Galitzen, *Vorlesungen über Seismometrie*, p. 138, 1914.

[2] *Physics of Earthquake Phenomena* (1908), chap. xii.

[3] *Op. cit.*, pp. 248–50.

[4] G. W. Walker, *Modern Seismology*, 1913, p. 61.

[5] *California Earthquake of April 18, 1906* (Report of the State Earthquake Investigation Commission, Vol. II, "The Mechanics of the Earthquake," by H. F. Reid). Published by the Carnegie Institution of Washington, 1910.

surface to 12.5 km. per second at 2,170 km. from the surface, 0.66 of the radius from the center. Below that depth the velocity is nearly constant. The velocity of the transverse waves is 4.8 km. per second at the surface and increases almost linearly with depth, reaching a velocity of about 7.5 km. per second at half the distance to the center of the earth. The absence of good records from distances beyond 125° prevents a knowledge of the velocities at greater depths. Within the limits regarding which information is given, Reid remarks that there is no indication of a sudden change in the velocity of either wave such as we should expect if there were any sudden changes in the nature of the earth's interior. Oldham also finds no evidence of sudden change to a depth of at least 2,400 miles, 0.4 radius from the center.[1] From the curves showing the relation of velocity to depth which Reid gives[2] it is seen that the ratio of velocity of the transverse to the velocity of the longitudinal wave is 0.66 at the surface, 0.56 at 0.95 R, 0.53 at 0.9 R, reaching a minimum of 0.52 at 0.85 R, from which it increases to 0.58 at 0.5 R. This shows that both moduli of elasticity increase with depth, but that down to a depth of between 0.8 and 0.9 R. from the center of the earth, 637 and 1,274 km. from the surface, incompressibility increases relatively faster than rigidity. The change is shown as very rapid in the first 300 km. This is the only way in which the existence of an asthenosphere reflects itself in the rigidity of the earth, and this may not be related to its weakness but to some other property, such as the nature of compressibility or of changing chemical composition, or partly in the lack of detailed knowledge in the nature of the data.

Earthquake waves, like the tides, measure elasticity rather than strength. The vibrations which penetrate 200–300 km., and more, downward in the earth are already greatly reduced in amplitude and therefore in the strains which they bring on the earth. What the maximum strains may be is unknown, but reasonable assumptions as to amplitude show that within the asthenosphere the order of magnitude of the strains would be of the nature

[1] "On the Constitution of the Interior of the Earth as Revealed by Earthquakes," *Quar. Jour. Geol. Soc.*, LXII (1906), p. 470.

[2] P. 122.

of a thousandth part of that which granite at the surface of the earth can sustain. Furthermore, even if the stresses were greater and could be used as a measure of strength, this would apply to sudden stresses only and the results obtained from elastic vibrations could not be used safely as a means of determining the strength under long-enduring stresses. Thus the evidence from both tides and earthquakes is negative in regard to the existence of an asthenosphere. They show only that it is not fluid and that it is not markedly unlike the rest of the earth in its elastic properties.

HIGH, BUT VARIABLE, ELASTIC LIMIT WITHIN THE UPPER LITHOSPHERE

The experiments by F. D. Adams showed that under conditions of cubic compression rocks became far stronger than when subjected to compression, as at the surface of the earth, in one direction only. When a cylinder of Westerly granite was incased in a steel jacket and then subjected to heavy pressure upon its ends, a small cavity within the specimen just began to break down under a stress-difference of between 160,000 and 200,000 pounds per square inch, about six to eight times the strength possessed by this rock under surface conditions. At a temperature of 550° C., a temperature calculated to exist at a depth of 11 miles below the earth's surface, small cavities remained open when submitted to considerably greater pressures than occur from the overlying load at this depth.[1]

Adams' experiments and King's calculations are most important and show without doubt that the more superficial parts of the earth, to a depth of ten to fifteen miles at least, are far stronger than had been supposed; but they apply to the temperature and pressure gradients in places of geologic quiet, not to regions undergoing igneous intrusion and crustal deformation. Then the temperatures may become far higher and the crust surcharged with magmatic gases. Yet it is under these conditions especially, of geologic activity as contrasted to geologic quiet, that regional metamorphism and rock flowage proceeds. Still less does this experimental work prove a great strength of the crust at depths of more

[1] Louis Vessot King, "On the Limiting Strength of Rocks under Conditions of Stress Existing in the Earth's Interior," *Jour. Geol.*, XX (1912), 136, 137.

than a hundred kilometers, for there the temperatures are presumably above those which under the conditions of freedom from pressure at the surface of the earth produce dry fusion. Occluded gases, furthermore, are held beyond possibility of escape.

The strength of the crust is dependent consequently upon fourfold conditions—the nature of the material, the cubic compression, the relation of temperature to the point of fusion, and the rapidity of the application of the stress. These factors are all variable with time and place. How variable will be seen upon further consideration in the following paragraphs.

The influence of the nature of the material is seen when it is noted that granite is only about one-half as rigid as the basic rocks, although it is not less strong. Consequently, regional stress coming upon a complex of two such rocks will elastically deform the granite more readily, a greater stress will be thrown upon the basic rocks, and since their elastic limit is not correspondingly higher they should begin to yield by flow or fracture before the more pliant rocks had reached their limit. The general conclusion is that a movement of compression in the earth's crust must necessarily give rise to unequal strains and concentration of stress, as well from variations in chemical composition as from variations in structure. The local stress may rise far higher than the general regional stress.

As to the second factor, during the progress of normal faulting the horizontal compressive stress in the crust is less than the vertical stress due to weight. During the progress of folding and mashing, on the contrary, the horizontal stresses become far higher. But the least of the three principal stresses determines the amount of cubic compression; the difference between the greatest and least stresses determines, on the contrary, the amount and direction of the strain upon the rigidity of the rock. Thus it is seen that both the cubic compression and the stress-difference vary with the amount and kind of forces.

It is temperature, however, which is probably the most variable of these factors. Igneous activity brings the temperatures of the greater depths comparatively near to the surface and must produce

widespread weakening of the crust, both through the physico-chemical effects of the exalted temperatures and the structural effects of the intruded viscous fluids.

The rapidity of the application of stress is a variable in itself and furthermore has variable effects, but would seem, however, to be the least important of these several factors. The movements of horizontal compression and vertical warping are slow and give time for recrystallization in the deeper crust. In this way they meet a lesser resistance than would rapid stresses. Where the temperatures are close to those of fusion it would seem in fact that rock flowage by recrystallization, developing the gneissoid structure, should demand markedly less shearing stress than the process of granulation. The gnarled and twisted rocks of the Archean speak of the presence beneath them of molten magmas rather than of an enormous degree of compressive forces upon them. But ready yielding by recrystallization in one place would permit the concentration of mashing stresses upon other localities and raise the strain to that intensity needed for granulation. An enormous depth of cover, such as Adams' experiments have been thought to show, is not suggested by the geologic evidence, nor apparently is it demanded by a completer theory.

In fault movements and in dike or sheet intrusion accompanied by the expansion of gases are two sources of rapid application of forces. It is probable, however, that their deformative action is confined to the outer ten miles of the crust, and their consideration need not detain us in the evaluation of those factors of strength which concern the crust as a whole.

Summing up the conclusions from these various lines of evidence, physical and geological, it is seen that they suggest a rapid increase of strength with depth, then the gradual passage into a deep zone of lowered strength. The limits and values, however, are variable with time and place. Such a distribution of strength as is indicated by these independent lines is in accord with the interpretation of the geodetic evidence showing the existence of crustal competence to support heavy loads over certain limits of area, coexisting with flotational equilibrium over much broader regions.

MODES OF LITHOSPHERIC YIELDING AND THEIR RELATION TO STRENGTH

The relationship of strength to depth which has been derived in this study and which was expressed in the curve of strength at the end of Part VII is to be connected with the physical qualities discussed in this part. Here it is seen that it is a curve of elastic limit. When that limit is exceeded, permanent deformation must take place; by one means at the surface, by another within the body of the lithosphere, by still another at its base.

At the surface the typical mode of yielding is by jointing and faulting, in stratified beds by folding also. The movements in this zone of fracture and in the transitional zone of combined fracture and flow may be regarded as merely the responses in a thin, brittle, and relatively weak outer layer to deformative movements progressing in the great thickness of the lithosphere below. But the rocks of deeper origin which have been exposed at the surface by profound erosion show that they have yielded in another fashion. Their foliated structures and crystalline textures testify to yielding by massive flowage. Fracturing appears to have been absent, except in so far as it was produced by intrusions from below, giving rise to complexes of dikes and sheets. These visible exposures suggest that at still greater depths, notwithstanding the great strength of that zone, open fracture planes disappear and rock flowage both by granulation and by recrystallization is still more distinctive. This appears then to be the mode of yielding of the great body of the lithosphere.

Recently Becker has suggested that fracturing may enter into the problem of isostasy in the following way: The demonstrated capacity of small cavities to remain open under great pressures may permit fissuring and jointing to extend deeper into the crust than had been previously thought possible. To the degree to which fractures and porosities do exist they must decrease the specific gravity of rocks. If shattering pervaded the rocks of one region and not another, even though the rocks were exactly alike in composition, the densities would become different. To give isostatic equilibrium the region of shattered rocks would have to stand higher than the other. This would be the initial effect as

a result of the decrease in density, even if the zone of compensation rested on an unyielding base.[1] The logical correctness of this argument is not to be questioned, but rather the degree of its application. The following arguments suggest that shattering or porosity are, however, very subordinate rather than determining factors in the isostatic problem.

Such a theory does not account readily for the movements needed to maintain isostasy because of erosion and sedimentation. These surface changes of mass suggest a restoration of mass by lateral undertow. Furthermore, the appeal to nature shows that the rocks, once deep-seated, which have become revealed at the surface by erosion, are almost without pore space. The average porosity according to Fuller is 0.2 per cent, but the mean differences in densities between ocean and continent which must be accounted for under the hypothesis of uniform compensation to a depth of 122 km. amount to about 4 per cent.

Joints are observed to decrease with depth, becoming tighter and more distantly spaced, and the indications given by the lack of general circulation of ground-water through crystalline rocks, except within joint spaces near the surface, are that at greater depth the joint spaces are negligible.

In the great compressive movements the whole thickness of the crust must yield, but even this cannot be conceived as producing porosity by granulation sufficient to notably modify the density. A large part of the deformation in the deeper crust must be by a process of recrystallization. Assume, however, that granulation is the dominant process. Observation of granulated rocks shows a reduction in size of the crystals, but these broken fragments fit against each other perfectly and without great internal distortion of crystals. In granulated rocks from the zone of flow there is therefore always some amount of recrystallization, sufficient to eliminate that porosity connected with minute shattering and movement of the broken particles. The explanation appears to be as follows: The minute shattering of the minerals tends to give a high pore space, but with a high pore space the amount of contact

[1] G. T. Becker, "Isostasy and Radioactivity," *Science*, XLI (1915), 157–60; "On the Earth Considered as a Heat Engine," *Proc. Nat. Acad. Sci.*, I (1915), 81–86.

between grains becomes proportionately less. For the prevention of ready recrystallization and the maintenance of this pore space the granulated rock, according to present theory, must be conceived of as dry and the grains accordingly unsupported except at the points of contact. The shear strains within each grain become very great in proportion to the diminution of contact, and increase in proportion to the regional pressure. If the points of contact, for example, cover only one-fourth of the surface, the compression on those points would be four times as great per unit of surface as if there were continuous contact between grains. On the intervening parts of the surface there would be no pressure. Internal shears would result in this way from the hydrostatic pressure of dry rock due to depth and are not dependent upon a pressure-difference in the rock as a whole. The internal strains would tend to produce molecular changes of state as in the plastic flow of metals. There would be melting to relieve the strain, and refreezing by which the molecules would build out the crystals into the pore spaces. By this means recrystallization can go on without the aid of crystallizers, though presumably with more difficulty, and the comminuted crystals come to fit compactly as they are observed to do. This elimination of porosity presumably goes on approximately with the process of granulation, though it may lag somewhat. It would go forward more effectively with depth, irrespective of temperature, since there would be the greater static load upon the rock and the greater differential pressures within the mineral particles. It might be expected that such reduction of pore space would go forward to a limited extent only, leaving a residual porosity. Observation, however, shows that the pore space has been almost completely eliminated. Furthermore, the rocks now exposed at the surface acquired their absence of pore space at depths of only a few miles from the surface. At depths measured in tens of miles there seems then no expectation that density would be notably decreased because of a development of porosity.

To sum up the modes of yielding within the lithosphere: at the surface is seen to exist a thin outer crust intimately cracked on the outside by closely spaced parallel joint systems. Local

extreme deformation is by faults and folds. With increasing depth and strength the joints become less abundant and faults pass into flexures. The passage of fractures into flexures implies the beginnings of massive flow. Where magmatic heat or emanations are not present the mode of mashing is presumably more especially by granulation. With still greater depth the yielding becomes more uniformly distributed throughout the rock mass. Both because of this pervasiveness of mashing and the great strength of this zone, deformation here requires the most force and absorbs the most energy of any part of the lithosphere. At greater depths the rock is more compressed, and is still more rigid than above, but the temperature here approaches fusion; recrystallization readily takes place, the strain which can be elastically carried is in consequence low, and the lithosphere passes gradually into the asthenosphere. Where, however, magmas rise through the crust they carry with them the environment of the asthenosphere; the lithosphere becomes locally abnormally heated and saturated with magmatic emanations. Recrystallization goes forward readily and the zone of weakness penetrates upward even to the zone of fracture. Thus in the injected and crystallized roofs of ancient batholiths, laid bare by profound erosion, we may perceive the nearest approach to dynamic conditions which prevail in depths forever hidden.

POPOSAURUS GRACILIS, A NEW REPTILE FROM THE TRIASSIC OF WYOMING

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Some time ago the writer mentioned a new form from the Triassic of Wyoming and briefly described one of the vertebra.[1] At that time it was hoped that the explorations in the western Trias of the past season would bring to light something further concerning this peculiar form. No additional evidence has so far been disclosed, however, in the collections from Arizona and New Mexico, and it is probably well at this time to describe more fully the remains for the benefit of others who may be engaged in the same field.

The material herein described consists of an ilium, several vertebrae from various parts of the column, two femora, a tibia, the proximal end of a fibula, and several innominate fragments. All are considerably crushed and broken. The leg bones are so flattened that it is rather difficult to determine the true proportions.

The ilium, as previously pointed out,[2] is very similar to that described by J. H. Lees[3] as belonging to *Paleorhinus Bransoni*. The condition of the bone suggests little distortion; still, upon a comparison with the specimen described by Lees, it is evident that the supra-acetabular flange has been somewhat depressed by pressure. To all appearances the acetabulum is formed almost entirely by the ilium. It is broad and deep and closely confined above by the supra-acetabular flange. In preparing the specimen what appears to be a foramen of perhaps 5 mm. in diameter was exposed, running in and slightly upward from the upper acetabular surface. It is barely possible, however, that this is simply a matrix-

[1] M. G. Mehl, "The Phytosauria of the Rocky Mountain Trias," *Jour. Geol.*, 1914 (in press).

[2] *Op. cit.*, p. 159.

[3] "The Skull of *Paleorhinus*, a Wyoming Phytosaur," *Jour. Geol.*, XV (1905), 44.

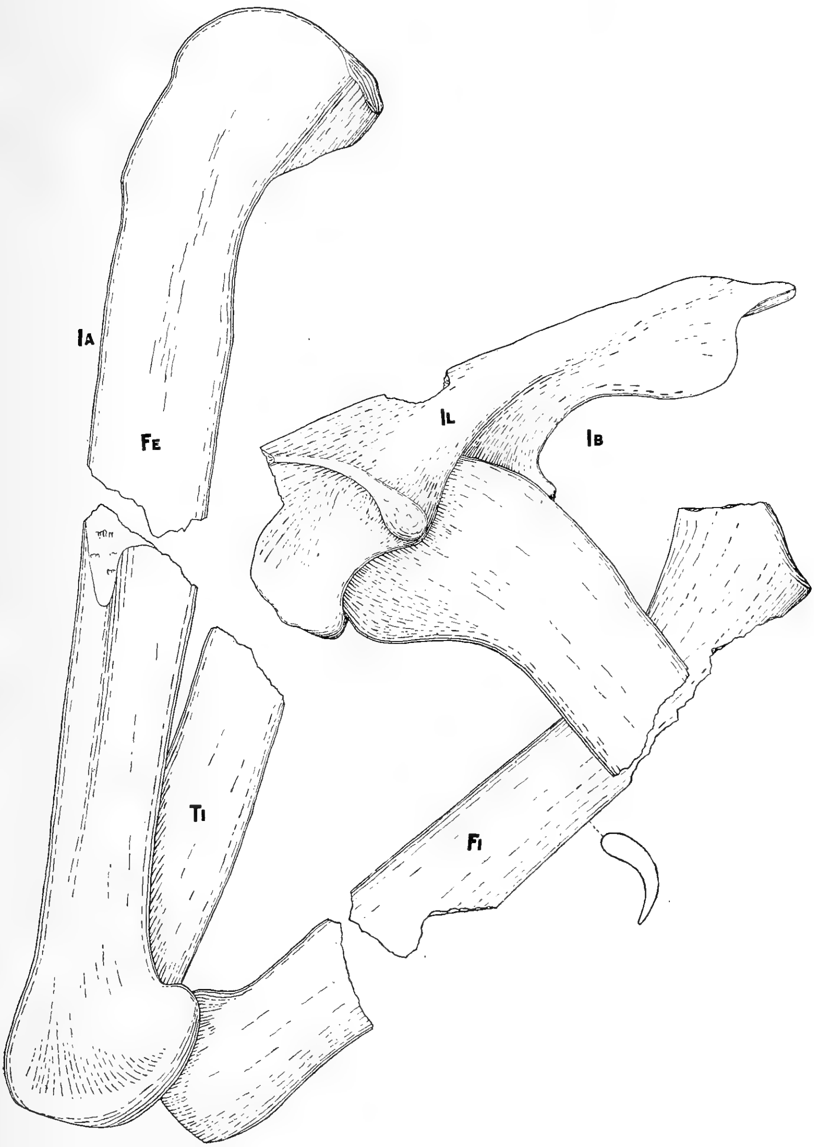

FIG. 1.—*Poposaurus gracilis.* 1*a*, left femur, tibia, and fibula as they lay in the matrix. 1*b*, right ilium and femur. All about $\frac{1}{3}$ natural size.

filled crack. From the center of the margin of the supra-acetabular flange a strong ridge runs up and back in a regular curve. Below, this ridge is about 20 mm. in width from which it gradually narrows till it becomes confluent with the upper posterior margin of the ilium. The upper posterior process is broken away, but the indications are that it was a thin, platelike process with rounded border, separated from the lower posterior process by a deep sinus. The upper anterior process is irregular in shape, extending forward in a regularly expanding spatulate form. A marked ridge extends along its outer surface from the supra-acetabular flange, becoming fainter anteriorly and finally confluent with the upper anterior border. The process makes up nearly half of the total length of the ilium, the greatest length of which is 250 mm. The lower anterior projection of the ilium has a somewhat antero-posterior elongated, convex face for the articulation of the pubis. The articular face for the ischium, the lower posterior process, is somewhat larger, triangular in outline and nearly flat. These two articular faces are separated by the thin, apparently complete lower border of the acetabulum, strongly suggesting a perforate acetabulum, as in the dinosaurs.

Of the dorsal vertebrae but two are sufficiently well preserved to be of use in description. They probably represent the anterior thoracics inasmuch as they show an early stage in the transition of capitular facet from the anterior face of the centrum to the neural arch. The description of one of these will suffice as they show essentially the same features.

The centrum is spool-shaped, moderately biconcave, and very much constricted laterally between the articular faces. This has probably been greatly accentuated by pressure. The articular faces are oval in outline, about 32 mm. wide and 39 mm. high. The centrum is 54 mm. long and was probably 15 mm. wide at the center in an uncrushed condition. The neural arch is high and delicately constructed. The diapophysis is a thin horizontal plate confluent anteriorly with the articular face of the zygapophysis and gradually expanding posteriorly. At its posterior extremity it suddenly thickens below for the tubercular facet which is separated from the posterior zygapophysis by a deep, rounded sinus.

At this point, the greatest width of the diapophyses, the two tubercular facets are approximately 50 mm. apart. Near the mid-length the diapophysis is supported below by two thin diverging buttresses, the anterior one confluent below with the upper anterior

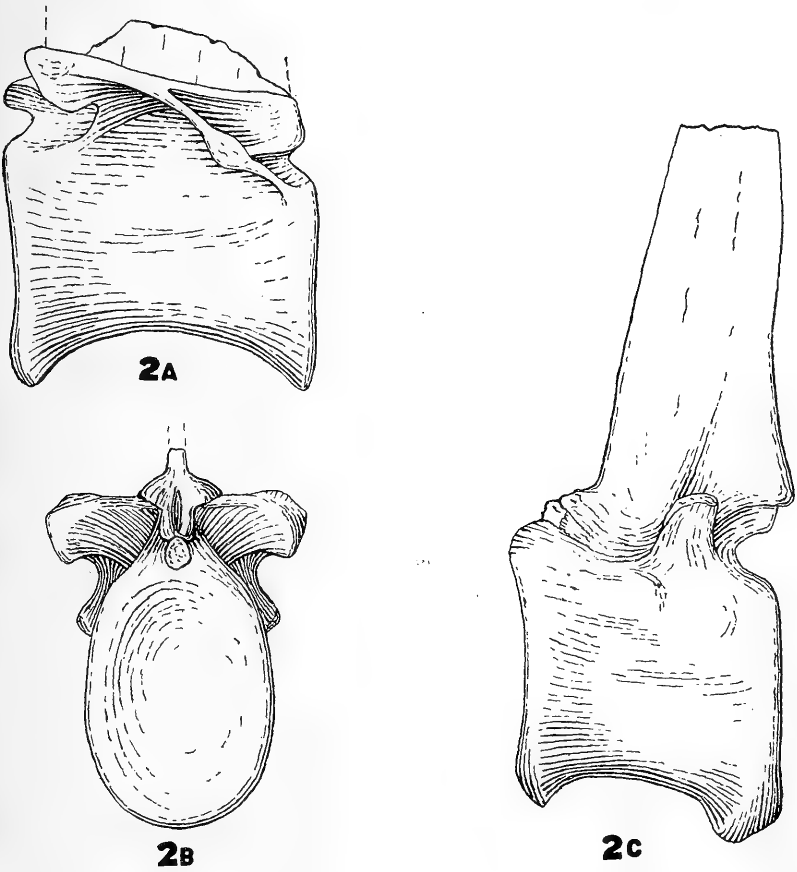

FIG. 2.—*Poposaurus gracilis.* *2a,* right lateral view of one of the dorsal vertebrae. *2b,* posterior view of the same. *2c,* left lateral view of second caudal vertebra. All about $\frac{2}{3}$ natural size.

face of the centrum. Just before reaching this point it is considerably swollen to form the capitular facet for the reception of the rib. The posterior support is directed toward the upper posterior face of the centrum, but loses its identity before reaching that

point. The distance between the capitular and tubercular facets is about 38 mm. The articular faces of the zygapophyses are of moderate size. The anterior pair are directed up and in; the posterior pair, of necessity, out and down. The anterior pair form a cuplike depression with a narrow antero-posteriorly directed median sinus into which apparently extended a thin vertical support of the posterior pair of zygapophyses of the preceding vertebrae, a condition that would apparently restrict somewhat the free motion of the vertebral column. But little remains of the spine. At the base it is 5 mm. thick and about 45 mm. wide.

The sacrum consists of at least four, and probably five closely united vertebrae with a total length of perhaps 230 to 290 mm. It is considerably abraded and lacks most of the neural arch. The centra of the component vertebrae rapidly decrease in size posteriorly. The anterior face of the first is about 45 mm. in diameter. All the centra are considerably constricted laterally. Apparently only the anterior two vertebrae bear sacral ribs, though this is not certain. The ribs were stout, somewhat flattened, and articulated low on the neural arch near the mid-length of the vertebra.

The caudal vertebrae are represented by the first and second of the series. The second is nearly perfect and will serve well to characterize the group. The centrum is 50 mm. long, somewhat more deeply biconcave than the dorsals, and much constricted laterally at mid-length. The articular faces are some 40 mm. wide and 50 mm. high. Below there is a strong inter-centrum articulation for the reception of chevron bones. These articulations are well developed on both the anterior and posterior faces, but considerably stronger on the anterior face of the vertebra. The diapophyses are placed high on the neural arch and well back. Their tips are slightly higher than the neural canal. They are short, directed upward, and slightly back. From tip to tip they measure 42 mm. The zygapophyses are placed close together, the anterior ones directed in and upward. The spine is long and thin and directed very slightly backward. At the base it is about 5 mm. thick and 35 mm. wide. The extremity is broken away, probably no considerable portion, leaving the total height of the vertebra 135 mm.

The first caudal differs from the second in that the centrum of the former is somewhat longer, has larger articular faces, and has the chevron articulation at the posterior end only. There is also a considerable difference in the diapophyses of the two vertebrae. In the first they are broad and, although badly weathered, suggest a lateral face of perhaps 30 mm. anterior-posterior extent as compared with 10 mm. in the second. In the first caudal the diapophyses are placed farther forward and slightly lower than in the second and are slightly supported by the centrum.

The condition of the leg bones is such that little can be done in the way of description other than stating approximate measurements. The femur is at least 465 mm. long and very slender. In an uncrushed condition the shaft was probably about 45 mm. in diameter. The head is considerably expanded antero-posteriorly, measuring 85 mm. in that direction. Little can be said of the condyles except that the antero-posterior extent of the articular face was great, perhaps 85 mm. or more.

A piece of the tibia, 125 mm. in length, a portion of the shaft near the proximal end, remains imbedded in the matrix in its natural relation with the left femur. In life the shaft was probably about 30 mm. in diameter. The left fibula, some 345 mm. in length, is very thin laterally and much expanded antero-posteriorly at both the proximal and distal ends. The shaft, which is 40 mm. in width, is regularly concavo-convex antero-posteriorly throughout its entire length, the convexity outward. Were this a feature due to distortion only, one would hardly expect the solid articular ends to conform to the curvature of the thin shaft as they do. Furthermore, polished cross-sections of the shaft do not suggest considerable flattening of its walls.

RELATIONSHIPS

The specimen has little in common with the phytosaurs; besides its apparently perforate acetabulum, of which there is but the slightest suggestion in some of the phytosaurs, it shows a marked difference in the shape of the ilium and in the comparatively large number of vertebrae fused together in the sacrum. In the phytosaurs there are but two sacrals and these are free. As *Dolichobrachium*, described by Dr. S. W. Williston from the Trias of

Wyoming,[1] is known from the humerus, teeth, and coraco-scapula only, there is little basis for comparison. The first two elements and especially the coraco-scapula are suggestive of a more massive build than that of the present specimen. There is, however, the possibility that these two forms are identical, an uncertainty that must remain till more abundant material is to be had. In some respects it resembles some of the early dinosaurs. The hollow, slender leg bones, the slightly biconcave vertebrae, are suggestive of the Theropoda and the perforate acetabulum is typically dinosaurian. Unlike the condition in this group of dinosaurs, however, each sacral rib is supported by a single vertebra. While this is a condition found in the sauropod dinosaurs, their solid bones, the expanded neural canal of the sacrum, and many minor points are very different from the form under discussion. In the future it may be shown that the genus *Palaeoctonus* of Cope is related to the form here described as, indeed, there is a possibility that *Palaeoctonus* and *Dolichobrachium* are identical. When the remains of the latter genus were discovered there were present portions of a skull in a powdery condition, too poorly preserved to be saved. According to Dr. Williston there was enough seen of the teeth, however, to suggest a possible relationship with *Palaeoctonus*.

Everything in the structure of the form so far studied indicates a well-muscled creature light in weight, possibly bipedal in gait occasionally, and most assuredly swift in movement. The name *Poposaurus gracilis* is suggested for the material herein described from its discovery on the Popo Agie river.

The type specimen is numbered 602 in the University of Chicago collections. It was collected by Professor E. B. Branson in the red beds near Lander, Wyoming.

[1] "Notice of Some New Reptiles from the Upper Trias of Wyoming," *Jour. Geol.*, XII (1904), 688.

THE CANNONBALL MARINE MEMBER OF THE LANCE FORMATION OF NORTH AND SOUTH DAKOTA AND ITS BEARING ON THE LANCE-LARAMIE PROBLEM[1]

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Field examination by the writers and the paleontological determinations by Drs. Stanton and Knowlton during the years 1912 and 1913 show that in a large region west of Missouri River in North and South Dakota the Lance formation consists of two distinct parts, a lower non-marine part containing a flora very similar to, if not identical with, that of the Fort Union and an upper marine member containing a fauna closely resembling, but not identical with, that of the Fox Hills sandstone. This upper part, on account of its peculiar fauna, has been mapped separately and named the Cannonball marine member of the Lance formation. Farther west non-marine beds bearing lignite and occupying a similar stratigraphic position have been named the Ludlow lignitic member of the Lance.

The area examined by the writers embraces a territory of over 5,000 square miles, extending from Mandan, North Dakota, west to Montana, a distance of about 175 miles (Fig. 1), and south beyond the boundary line of South Dakota. The marine member of the Lance has been mapped from Mandan to a point 4 miles west of Haley, North Dakota, a total distance of about 130 miles. A large part of the area examined by the writers has been described already from an economic standpoint.[2]

The Lance formation had previously been mapped in two adjoining regions, the Bismark Quadrangle[3] and the Standing Rock

[1] Published by permission of the Director of the U.S. Geological Survey.

[2] E. Russell Lloyd, "The Cannonball River Lignite Field, Morton, Adams, and Hettinger Counties, North Dakota," *U.S. Geol. Survey Bull. 541–g;* D. E. Winchester, C. J. Hares, E. M. Parks, and E. Russell Lloyd, "The Lignite Field of Northwestern South Dakota," *U.S. Geol. Survey Bull.* (in course of preparation); C. J. Hares, "Lignite in Southwest North Dakota," *U.S. Geol. Survey Bull.* (in course of preparation).

[3] A. G. Leonard, *U.S. Geol. Survey Geol. Atlas*, Bismarck Folio (No. 181), 1912.

Indian Reservation,[1] in both of which one of the writers (Lloyd) has been able to distinguish the upper marine member. Parts of these areas are included in the accompanying map (Fig. 1).

Throughout the greater part of the area studied by the writers the strata are nearly flat-lying, with only a very low general dip, usually less than 30 feet per mile, to the north or northeast. Practically the entire area is grass-covered and only in the bluffs of the larger streams and locally in high steep-sided buttes are natural rock exposures found. The relief is very slight; hence there is considerable difficulty in tracing the formation boundaries.

The general character and relations of the Cretaceous and Tertiary formations exposed in this part of the Dakotas are shown in Table I. The areal distribution is shown on the map (Fig. 1).

TABLE I

CRETACEOUS AND TERTIARY FORMATIONS IN WESTERN NORTH AND SOUTH DAKOTA

System	Series	Formation	Character	Thickness in Feet
Tertiary	Oligocene	White River	Cross-bedded sandstone and fresh-water limestone	140
Unconformity, angular up to 20° and erosional up to 1,500 feet				
Tertiary	Eocene	Fort Union	Yellow sandstone, shale, clay, and lignite	1,025
Tertiary?	Eocene?	Lance	Cannonball marine member. Dark sandy shale, dark shaly sandstone, and yellow sandstone, containing marine shells	0–300
			Ludlow lignitic member. Light sandy shale, calcareous sandstone and lignite	0–350
			Somber-colored shale, yellow sandstone, and thin lignite beds	400–525
Cretaceous	Upper Cretaceous	Fox Hills	Sandstone, yellow	25–400
		Pierre	Dark marine shale	300+

[1] W. R. Calvert and Others, "Geology of the Standing Rock and Cheyenne River Indian Reservations, North and South Dakota," *U.S. Geol. Survey Bull. 575*, 1914.

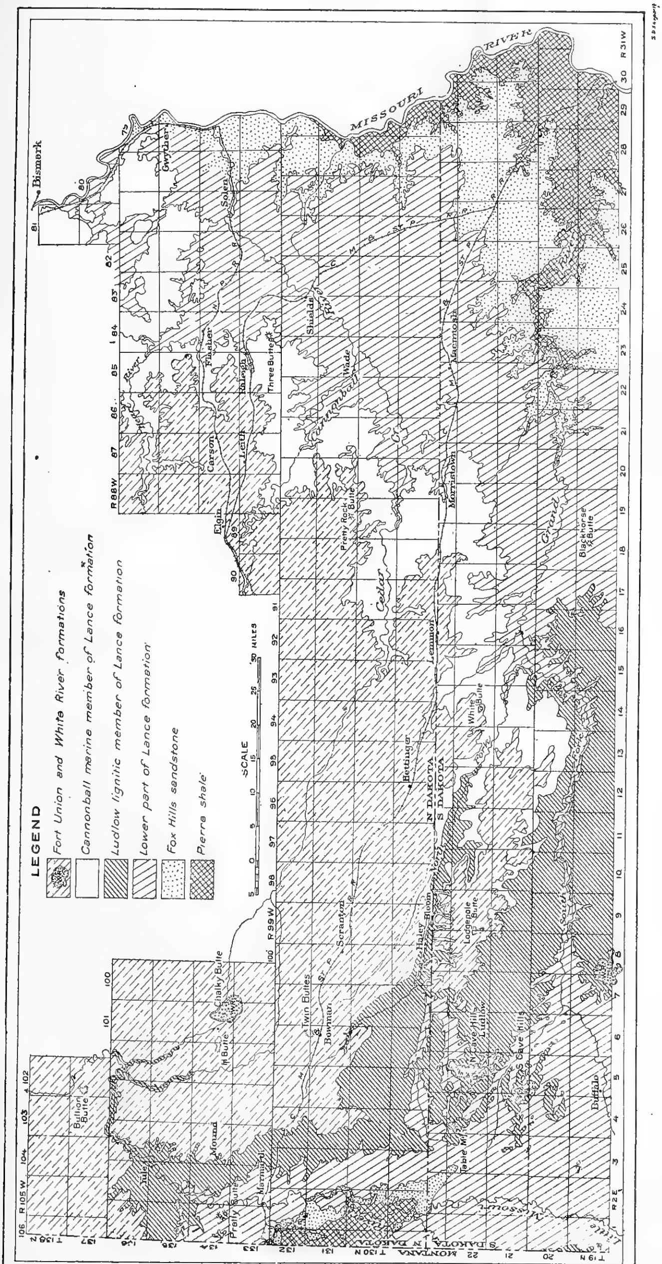

FIG. 1.—Geologic map of part of western North and South Dakota. Compiled by E. Russell Lloyd and C. J. Hares.

CRETACEOUS SYSTEM

FOX HILLS SANDSTONE

The type locality of the Fox Hills sandstone is Fox Ridge in the Cheyenne River Indian Reservation in South Dakota between Moreau and Cheyenne rivers and about 25 miles northwest of the mouth of the latter stream.[1] This region was examined in detail by V. H. Barnett in 1909, and the formation was mapped by Barnett and other geologists of the U.S. Geological Survey northward to Cannonball River.[2] The Fox Hills is made up of massive gray, yellow, and buff sandstone and banded shale, the maximum thickness in the Standing Rock and Cheyenne River Indian reservations being 400 feet and the average about 150 feet. Around the southern end of the Glendive anticline in Billings and Bowman counties, North Dakota, the Fox Hills sandstone is about 75 feet thick. It has been shown that in some localities in the Dakotas and eastern Montana[3] the overlying Lance formation is unconformable on the Fox Hills, the contact between the two being marked by erosion channels in the Fox Hills sandstone. In one locality in the Standing Rock Reservation the Fox Hills is 25 feet or less in thickness, so that either the erosion interval represented is of considerable magnitude or else the formation is peculiarly variable in thickness. This evidence of unconformity is, however, vitiated by the presence in the Standing Rock Reservation of a bed above the line of unconformity containing brackish water and marine fossils and along Little Missouri River by the gradation by alternation of beds of Fox Hills character with those of Lance character. The evidence both for and against the supposition of an unconformity of major importance at this horizon is fully discussed in the papers cited. From the stratigraphic stand-

[1] The name was first used by Meek and Hayden, *Acad. Nat. Sci. Philadelphia, Proc.*, XIII (1862) 419.

[2] W. R. Calvert and Others, *op. cit.*; T. W. Stanton, "Fox Hills Sandstone and Lance Formation ('Ceratops Beds') in South Dakota, North Dakota, and Eastern Wyoming," *Am. Jour. Sci.*, 4th ser., XXX (1910) 172, 188; F. H. Knowlton, "Further Data on the Stratigraphic Position of the Lance Formation ('Ceratops Beds')"; *Jour. Geol.* XIX (1911) 358–74.

[3] F. H. Knowlton, *op. cit.*

point it appears to the writers that the local unconformities that have been observed are such as would be expected where a marine formation is succeeded by one of continental origin.

A significant feature of the Fox Hills sandstone and underlying Pierre shale in the Standing Rock and Cheyenne River Indian reservations is that the fauna found in the sandstone cannot be distinguished from that found in the upper part of the shale, so that the strata from which the Fox Hills fauna was collected include not only the Fox Hills sandstone but also about 200 feet of the underlying Pierre shale.

TERTIARY? SYSTEM

LANCE FORMATION

Lower part.—It has been previously stated that the Lance formation in this region consists of three parts, a lower group of shale and sandstone beds of continental origin and two contemporaneous upper members, one of sandstone and shale of marine origin, the other of sandstone, shale, and lignite of non-marine origin. The lower part of the formation outcrops in a wide belt of country in Morton County, North Dakota. Northward it passes below the flood plain of Missouri River a few miles below Bismarck, and to the west it has been mapped to and beyond the Montana-Dakota state line. It occupies a large area in Bowman and Billings counties in southwest North Dakota and in adjacent parts of Montana. Throughout this whole region it is essentially uniform in character, consisting predominantly of somber-colored arenaceous shale intercalated with lenticular beds of brown or buff sandstone. Beds of brown carbonaceous shale, bog iron ore, and thin lignite are conspicuous in most outcrops. All the strata of the lower part of the Lance are lenticular in character, and a section exposed at one locality is different in most of its details from one even a short distance away. Cross-bedding is common, especially in the sandstone. Irregularity of deposition is characteristic. Near Solen on Cannonball River, 9 miles above its mouth, the lower part of the Lance has a thickness of approximately 400 feet and on Little Missouri a thickness of about 525 feet.

The fossils found in the lower part of the Lance in the western Dakotas consist of plant impressions and bones of reptiles. The plant remains are not abundant, but a few good collections have been made from widely separated localities. All of the species belong to the widespread Fort Union flora. The vertebrates include turtles and several genera of dinosaurs, among which is the large-horned Triceratops, which is diagnostic of the Lance formation. The Ceratopsian remains are especially abundant near the base of the formation in the Standing Rock Indian Reservation, but on Little Missouri River several specimens of Triceratops have been found near the top of the formation and stratigraphically below the *Ostrea glabra* zone.

Ludlow lignitic member.—The Ludlow lignitic member of the Lance formation occupies a large area in Harding County, South Dakota, and has been mapped northward into Bowman and Billings counties, North Dakota, and eastward into Perkins County, South Dakota, where it merges with the Cannonball marine member. In the vicinity of Ludlow, South Dakota, its type locality, it consists of 350 feet of loosely consolidated buff and cream-colored calcareous sandstone and shale with interbedded lignite. It contains most of the lignite of South Dakota and the presence of this lignite is one of the chief criteria for considering it a distinct member of the Lance formation. Its lithologic character in South Dakota is very like, and its fossil flora so far as determined is identical with, the Fort Union. Its flora is like that of the lower part of the Lance, but its lithology is quite different. On the other hand, in North Dakota its flora has the same affinities as in South Dakota, but lithologically it resembles the lower part of the Lance, except for the presence of the numerous lignite beds. It is this variation in color and lithology of the Lance that renders its separation from the overlying Fort Union so difficult.

The following sections show the lithologic character of the Ludlow lignitic member.

Composite Section in Secs. 32 and 36, Tp. 22 N., R. 5 E., Black Hills Meridian, South Dakota

	Ft.	In.
Fort Union formation		
Sandstone, yellowish, and shale	255	0
Ludlow lignitic member of the Lance formation		
Lignite		8
Sandstone, somewhat shaly	45	0
Shale, dark	2	0
Lignite	2	4
Sandstone, light colored, grayish, argillaceous	12	0
Sandstone, buff, with ferruginous specks	15	0
Lignite		2
Sandstone, buff, fine-grained, muscovitic	11	0
Liginite		3
Sandstone	8	0
Lignite		6
Sandstone, buff	25	0
Shale, bluish	1	0
Shale, carbonaceous		10
Shale, arenaceous	5	0
Lignite		1
Sandstone, buff, soft	5	4
Sandstone	1	0
Shale, arenaceous with carbonaceous streaks	4	0
Sandstone, argillaceous	6	0
Lignite	1	0
Concealed interval	22	0
Sandstone, drab, cross-bedded, ripple-marked	3	0
Shale, arenaceous	4	0
Lignite	1	5
Shale, brown	3	0
Lignite	2	1
Shale, dark		5
Lignite		8
Shale	1	3
Lignite	1	5
Shale, brown	4	0
Lignite	2	3
	446	8

Partial Composite Section, Tp. 20 N., R. 9 E., Black Hills Meridian, South Dakota

	Ft.	In.
Ludlow lignitic number of the Lance formation		
Sandstone, yellow, medium fine-grained, capping hills....	20	0
Shale, brown..................................	3	0
Lignite.......................................	1	0
Shale, brown..................................		6
Sandstone, yellow.............................	16	0
Shale, brown..................................	1	10
Lignite, dirty................................		8
Shale, brown..................................	3	0
Shale, black..................................		8
Lignite, dirty................................		6
Shale, brown..................................	1	6
Sandstone, buff to yellow.....................	12	0
Shale, brown..................................	2	0
Shale, black..................................		4
Lignite, fair.................................		8
Sandstone and shale, with lignite streaks	13	6
Lignite, fair.................................		8
Shale...		3
Lignite, good.................................	3	6
Shale, brown, arenaceous......................	1	0
Sandstone.....................................		
	82	7

Composite Section in Tp. 21 N., R. 7 E., Black Hills Meridian, South Dakota

	Ft.	In.
Fort Union formation		
Sandstone and shale...........................	138	0
Ludlow lignitic member of the Lance formation		
Shale, chocolate-colored......................	2	0
Sandstone, buff, yellow, fine-grained, alternating with buff shale....................................	9	0
Shale, chocolate-colored......................	3	4
Lignite	1	0
Shale, buff and yellow sandstone, alternating............	16	0
Shale, chocolate-colored......................	3	0
Sandstone, brown, fine-grained................	2	6
Shale...	10	0
Lignite, Gianniatti bed.......................	11	8
Sandstone, light, argillaceous................	16	4

	Ft.	In.
Shale, brown	3	0
Lignite	4	2
Shale	13	0
Lignite	3	8
Shale and sandstone	15	0
Lignite, poor	3	4
Shale, carbonaceous		6
Shale		
	255	6

SECTION ON WEST SIDE OF LITTLE MISSOURI RIVER, SEC. 10, TP. 135 N., R. 105 W., NORTH DAKOTA[1]

	Ft.	In
Fort Union Formation		
Sandstone, light	15	2
Ludlow lignitic member of Lance formation		
Shale, dark	24	0
Limestone, arenaceous	2	10
Shale, blue	7	2
Sandstone, gray, friable	4	0
Shale, gray	3	7
Sandstone, argillaceous	1	4
Shale, drab, hard	1	2
Lignite	4	5
Shale, dark, hard	2	10
Shale, light, fossils, leaves	2	10
Lignite	2	6
Shale, carbonaceous	1	8
Sandstone	2	8
Shale	1	8
Sandstone and shale, alternately	2	8
Lignite	1	0
Sandstone and shale	3	0
Lignite		6
Sandstone	4	0
Lignite	1	7
Sandstone, white, hard	3	0
Sandstone in upper part, shale in lower part; *Ostrea glabra* above base (only representative found of Cannonball marine member)	33	11

[1] See footnote, p. 533.

	Ft.	In.
Lignite	1	11
Sandstone and shale, sandstone in upper part and shale below	27	0
Lignite		10
Shale	11	0
Lignite	3	0
Shale	3	4
Sandstone, hard ledge in upper part	13	2
Lignite	5	3
Shale	1	8
Lignite	1	2
Shale	1	4
Lignite	5	0
Shale	8	6
Lignite	6	11
Shale	4	8
Lignite		6
Shale	7	6
Lignite	5	4
Sandstone	15	4
Shale		
	250	11

Section Along North Fork of Grand River at Bloom, South Dakota[1]

	Ft.	In.
Fort Union formation(?)		
1. Sandstone, brown and yellow, fine-grained, thin-bedded, interbedded with lenses of compact, bluish-gray limestone		
Cannonball marine member of the Lance formation		
2. Sandstone, dark gray, very calcareous, marine fossils *Nucula* sp., *Callista deweyi*, *Pholas* sp., *Corbula* sp., *Anchura americana*	10	0
Ludlow lignitic member of the Lance formation		
3. Lignite	2	0
4. Shale and sandstone, interbedded	40	0
5. Lignite	1	0
	53	0

Note.—No. 1 was considered as Fort Union but very possibly is Cannonball, as indicated by more recent field work in other parts of the field.

[1] See footnote, p. 533.

Generalized Section along North Fork of Grand River between Bloom, South Dakota, and Haley, North Dakota[1]

	Ft.	In.
1. Sandstone, brown and yellow, fine-grained, thin-bedded, and lenses of limestone	30	0
2. Sandstone, marine; burrows of a Teredo-like shell, *Anchura americana*	10	0
3. Shale	1	0
4. Lignite	2	0
5. Shale and sandstone, interbedded. Sandstone in lower part, light brown, medium fine-grained, micaceous; containing *Halymenites major*, *Thracia* sp., *Nucula* sp., *Callista* sp., *Pholas* sp., imprints of pelecypods and shark teeth	40	0
6. Lignite	1	0
7. Shale	4	0
8. Lignite	2	3
9. Shale		
	90	3

Note.—No. 1 was considered Fort Union but may be Cannonball; Nos. 2 and part of 5 are marine, and Nos. 3, 4, 6, 7, 8, and 9 are considered fresh-water.

Cannonball marine member.—The Cannonball marine member of the Lance formation has been traced and mapped in an area extending from near Mandan, North Dakota, to the vicinity of Ralph post-office in Harding County, South Dakota, and 4 miles west of Haley, Bowman County, North Dakota. The presence of brackish-water fossils, *Ostrea glabra*, near Yule on Little Missouri River in Billings County, North Dakota, shows that the sea probably extended some distance farther to the west than its sediments have been mapped. The extent of this member east of Missouri River is unknown.

The marine member is composed predominantly of dark sandy shale or shaly sandstone with a subordinate amount of dark-yellow and gray sandstone. It also contains some thin limestones. All the strata are lenticular in character and individual beds can be followed for only short distances. The member is typically exposed in the bluffs of Cannonball River in Tps. 132 and 133 N., Rs. 87 and 88 W. The following sections measured in the type area show

[1] The last three sections expose the interrelations of the Ludlow lignitic and the Cannonball marine members.

the character of the strata of the Cannonball marine member and of the strata which underlie and overlie it.

Section near Base of the Cannonball Marine Member, in South Bluff of Cannonball River, Sec. 33, Tp. 132 N., R. 87 W.

	Ft.	In.
1. Sandstone and sandy shale, yellowish gray	23	0
2. Sandstone, yellowish gray, unconsolidated, with thin lenses of shale near the bottom, containing poorly preserved leaf impressions	22	0
3. Shale, brown, sandy	22	0
4. Shale, bluish brown	2	0
5. Lignite		4
6. Shale, bluish brown	4	0
7. Sandstone, gray, unconsolidated	45+	0
8. Concealed to river level	15	0
	133	4+

In this section no definite line could be drawn between the lower part of the Lance and the Cannonball member. No. 1 and probably No. 2 belong to the marine member.

Section in South Bluff of Cannonball River, about Three-fourths of a Mile West of Foregoing Section

	Ft.	In.
1. Sandstone, yellow	25	0
2. Shale, containing poorly preserved plant remains	4	0
3. Sandstone, gray		
	29	0

The sandstone No. 1 in this section appears to be unconformable on the shale. It is probably the lowest bed of the Cannonball member.

Section in Bluffs of Cannonball River, near Janesburg, North Dakota, Tp. 132 N., R. 88 W.

	Ft.	In.
1. Sandstone, fine-grained, containing "cannonball" concretions	40±	0
2. Shale, dark, sandy	10	0
3. Sandstone and sandy shale, light colored, with "cannonball" concretions	21	6
4. Shale, dark, sandy, with *Perna* n.sp.	4	0

	Ft.	In.
5. Sandstone and shale, light colored, thin-bedded, the shale containing indeterminable vegetable remains	30	0
6. Sandstone, consolidated	1	6
7. Shale, light blue	1	6
8. Sandstone and shale, alternating	15	0
9. Shale, dark, carbonaceous	1	0
10. Sandstone and shale, alternating, base concealed by talus at base of slope	5+	
	129	6+

No. 4 and the overlying beds of this section are undoubtedly of marine origin; the lowest beds may be non-marine but no definite line could be drawn.

Section of Part of Cannonball Marine Member in South Bluffs of Cannonball River, E. ½ Sec. 5, Tp. 132 N., R. 88 W.

	Ft.	In.
Shale, sandy, black and brown, with *Pyrifusus newberryi* (?), fragments of *Crassitellites*, and a coral belonging to an undetermined genus	16	6
Sandstone, yellow, largely consolidated, but with lenticular concretions of hard sandstone	31	6
Shale, sandy, dark gray, nearly black	52	0
	100	0

Section of Lower Part of Fort Union Formation and Upper Part of Cannonball Marine Member of Lance Formation in South Bluff of Cannonball River, Secs. 31 and 32, Tp. 133 N., R. 88 W.

	Ft.	In.
Sandstone, yellow. Upper part massive, lower part thin-bedded and near base interbedded with thin bands of gray shale	30	6
Sandstone, dark gray to black	3	6
Sandstone, yellow, thin-bedded, interbedded with gray shale	13	6
Sandstone, gray, concretionary		6
Shale, bluish gray and somber-sandy	52	0
Shale, dark gray	54	6
Sandstone, yellowish gray, unconsolidated	5	0
Sandstone, yellow, hard at top, containing *Lunatia concinna*, *Turris minor*, *Fasciolaria buccinoides*, *Pyrifusus* sp., *Anchura* sp., and an undescribed coral	6	0
Shale, sandy, dark gray, grading at top into sandstone	31	6
	197	0

The upper sandstone of this section probably belongs to the Fort Union formation and the remainder to the Cannonball member of the Lance formation.

Sections similar in general character to those given above are exposed in the bluffs of Heart River and on Cedar Creek. The general appearance of the more sandy rocks is very much like that of the Fox Hills sandstone. The shale is very much like the Pierre.

A peculiar feature of both the Fox Hills sandstone and the Cannonball member of the Lance is the abundance of round concretions commonly known as "cannonballs." They are formed by cementation of the sandy shale by the deposition of calcium carbonate. These are true septarian nodules with radiating and concentric veins of calcite. The best examples observed are in a railway cut a few miles west of Raleigh, North Dakota. Where the "cannonballs" are exposed by stream erosion they are mostly weathered and broken to pieces.

The rocks of the Cannonball member weather typically into rounded hills, and in the interstream areas natural rock exposures are very few. In a large part of the area where the member was examined in South Dakota the only evidence that the beds are of marine origin are a few thin beds of nodular fossiliferous limestone, the presence of which is shown by the lines of residual bowlders at the borders of level-topped hills. The topography is such, however, that these horizons can be followed for long distances. No definite line could be drawn in the field between the Cannonball marine member and the lower part of the Lance. The contact of the two groups of strata is exposed at only a few places and in all such cases it seems to be impossible to tell where the beds of non-marine origin stop and those of marine origin begin. It follows from the foregoing statement that there is no evidence of unconformity at this horizon.

The relationship of the Cannonball member with the underlying lower part of the Lance is well shown in the region northwest of Solen on Cannonball River, North Dakota. In this area the characteristic chaos of badlands, formed by the erosion of the sandstone and shale of the lower part of the Lance, is bordered on

the north by a high, nearly level plateau which is capped by the lower fossiliferous sandstone of the Cannonball member. The top of the plateau is approximately 500 feet above and only about 2 miles distant from the river. The characteristic badland-forming strata of the lower part of the Lance extend up to within about 100 feet of the top of the plateau but here, as elsewhere, no sharp line of demarkation can be drawn.

Dr. Stanton has kindly prepared the following statement concerning the fauna of the Cannonball member.

The fauna of the Cannonball marine member of the Lance may now be characterized as a modified Fox Hills fauna. It contains a considerable proportion of undescribed species of Cretaceous affinities, and it is noteworthy that a number of the most common Fox Hills species have not been discovered in this fauna. The list of forms recognized is as follows:

Nodosaria sp.
Caryophyllia ? sp.
Anomia sp.
Perna sp.
Crenella sp.
*Cucullaea shumardi M. and H.
†Glycimeris subimbricata (M. and H.)
*Leda (Yoldia) scitula M. and H.
*Leda equilateralis M. and H. ?
*Nucula planimarginata M. and H.
‡Crassatellites evansi (H. and M.)
Solemya ? sp.
*‡Lucina occidentalis (Morton)
Corbicula cytheriformis M. and H.
†Cyprina ovata M. and H.
*Cyprina ovata var. compressa M. and H. ?
Veniella ? sp.
Callista sp. *a*
Callista sp. *b*
Tellina ? sp.
Thracia sp. related to T. subgracilis Whitfield
†Teredo globosa M. and H.
†Teredo selliformis M. and H.
Corbula sp.
Entalis sp.
Scala ? sp.
Turritella ? sp.
‡*Lunatia concinna (H. and M.)
Cerithium ? sp.
‡*Anchura americana (E. and S.)
Anchura americana (E. and S.) robust variety
Helicaulax ? sp.
†*Cantharus (Cantharulus) vaughani M. and H.
*Pyrifusus (Neptunella) newberryi M. and H. ?
*Fasciolaria buccinoides M. and H.
*Fasciolaria (Piestochilus) culbertsoni M. and H.
*Turris contortus M. and H.
Turris sp. related to T. contortus M. and H.
*Turris minor (E. and S.) ?
Cinulia sp.
*Cylichna scitula M. and H. ?

In this list of about 40 forms there are 21 named species and varieties of which 15 (marked *) occur in the Fox Hills, 4 (marked ‡) occur in the Pierre, and 5 (marked †) were originally described from rocks now known to belong to the marine member of the Lance. One species, *Corbicula cytheriformis*, was described from the Judith River formation and is known in the Mesaverde

formation and the Lance of other areas. The fossil seaweed, *Halymenites major*, which is common in the Fox Hills and other sandy formations of the marine Cretaceous, is also associated with the above listed fauna.

FORT UNION FORMATION

West of Missouri River Fort Union formation overlies the Cannonball member of the Lance and on Little Missouri it overlies the Ludlow lignitic member of the Lance. In this part of the Dakotas a large part of the formation has been removed by erosion, but in Billings County, North Dakota, one of the writers (Hares) has found a thickness of 1,025 feet. The formation consists of calcareous sandstone and shale of continental origin, containing numerous thick persistent beds of lignite and an abundant flora and fresh-water invertebrate fauna. Some of the thick lignite beds have burned extensively, and great numbers of red hills composed of fused and baked rock are characteristic features of the formation. The lower 100 feet of the formation is made up almost wholly of partially consolidated yellow and gray fine-grained sandstone which is in some localities indistinguishable from the sandstone at the top of the Cannonball member. At numerous other exposures, however, where the upper bed of the Cannonball member is a sandy shale, the contact is easily followed. At one locality on the north bank of Heart River in Tp. 136 N., R. 88 W., there is an erosion channel from 30 to 50 feet deep in the Cannonball member filled with the channel deposits of a Fort Union stream. Similar channel sandstones at the base of the Fort Union were seen along Little Missouri, near Yule, and at the mouth of Deep Creek, and also on Sand Creek, in Billings County, North Dakota. These strata show evidence of a rapid change in the character of the sedimentation. They consist predominantly of coarse sandstone containing lenses or pockets of conglomerate and of soft clay shale. The conglomerate consists of pebbles derived from the strata of the surrounding region and contain numerous water-worn bones, teeth, fish scales, fragments of silicified wood, and lignite in the form of tree trunks. Among the vertebrate remains are two mammalian teeth which have been identified by Dr. J. W. Gidley of the U.S. National Museum as *Euprotogonia* sp., second lower molar of left jaw, and *Pantolamda cavirictus*, upper premolar. Both of these species are found in the Fort Union beds of Sweetgrass

County, Montana. The remaining fragments are identified as representing *Champsosaurus*, a crocodile, a turtle, and *Lepisosteus*.

Lenses of conglomerate similar to the one described above were seen at other localities near the base of the Fort Union along Heart River, and in one exposure a bed of conglomerate a few inches thick forms the base of the formation. Along Little Missouri in sec. 31, Tp. 138 N., R. 102 W., a conglomerate at the base of the Fort Union contains bowlders up to a foot in diameter. The unconformity shown in these exposures is such as would be expected in a transition from marine to continental sedimentation.

ROCKS OF POST-FORT UNION AGE

The rocks of post-Fort Union age in this region embrace: (1) small remnants of sandstone, marl, and limestone of the White River formation on the tops of a few high buttes; (2) sand and gravel beds on the tops of high buttes, probably deposited by streams previous to the present cycle of erosion and derived in part from the White River formation; (3) terrace gravel in the valleys of the larger streams from 50 to 250 feet above the present valley floors; and (4) a large number of scattered glacial bowlders, the remnants of the drift of one of the earlier glacial epochs. These later rocks occupy small areas and are not shown on the accompanying map.

INTERRELATIONSHIPS OF THE LUDLOW LIGNITIC AND THE CANNONBALL MARINE MEMBERS OF THE LANCE FORMATION

It has been shown that the Lance formation in a large region immediately west of Missouri River consists of two parts, the upper of which, the Cannonball member, is marine and contains a fauna similar to, but not identical with, that of the Fox Hills sandstone. The Cannonball member becomes gradually thinner toward the west, and the sea in which it was deposited perhaps did not extend as far west as the Montana line. The oyster beds near Yule in Billings County, North Dakota, first discovered by Leonard and later described by Stanton,[1] may represent the western-

[1] T. W. Stanton, "The Age and Stratigraphic Relations of the 'Ceratops Beds' of Wyoming and Montana," *Washington Acad. Sci., Proc.*, XI, (1909), 249; "Fox Hills Sandstone and Lance Formation ('Ceratops Beds') in South Dakota and Eastern Wyoming, *Am. Jour. Sci.*, 4th ser., XXX (1910), 183–84.

most limit to which the Cannonball sea extended. This region was studied in 1911, and it was found by Hares that the beds containing the oysters are about 700 feet above the base of the Lance.

The zone in which *Ostrea glabra* and *Ostrea subtrigonalis* occur, in Tps. 136, 134, and 134 N., R. 105 W., is considered the westward extension (so far as known at present) of the strata of marine origin, as the oysters are brackish-water animals and consequently must have had some connection with the open sea. The most westerly collection of the Cannonball marine fauna is only 30 miles east of this place, in sec. 21, Tp. 129 N., R. 100 W., 4 miles west of Haley, and occurs stratigraphically within 100 feet of the T Cross lignite bed which was traced from Tp. 134 N., R. 105 W., to the west side of Tp. 129 N., R. 101 W. The oysters occur about 70 feet above the same lignite bed; it is assumed that the seaward connection was to the east. The oysters also occur about 120 feet below the base of the Fort Union formation which in the Little Missouri region has the same characteristics (channel conglomerate, light-yellow, somewhat massive sandstone) as it does in the Cannonball River country where it rests directly on the Cannonball marine member. It appears that the sea in which were deposited some 300 feet of marine sediments transgressed westward across the lignitic strata of the Ludlow member and that the position of its westward limit is underneath the divide between the drainage of Little Missouri and that of Grand and Cannonball rivers. All of the Triceratops collected in the Little Missouri country came from below the T Cross lignite bed and the oysters from above it. Calvert, however, states that in Montana "Ceratopsian bones were found just above the lowest persistent lignite bed, but there is certainly nothing in the character of the overlying strata to suggest that similar bones do not occur therein up through a stratigraphic distance of perhaps 500 feet."[1] The T Cross lignite bed was mapped to the Montana state line and it is undoubtedly the same lignite as the "persistent lignite" referred to above.

The presence of numerous lignite beds in the upper part of the Ludlow lignitic member of the Lance is in strong contrast to the

[1] W. R. Calvert and Others, "Lignite in Eastern Montana," *U.S. Geol. Survey Bull. 471*, p. 197, 1912.

undifferentiated lower part of the formation, and the absence of marine fossils in this member is in contrast to the Cannonball marine member. The Ludlow lignitic and the Cannonball marine members are considered to be contemporaneous in age.

RELATIONSHIPS IN OTHER AREAS

The Cannonball sea presumably advanced into western North and South Dakota from the east or northeast, and by inference the Cannonball member continued with undiminished or with increased thickness to the north and northeast. Practically the whole region east of Missouri River is deeply covered with a mantle of the drift of the last Wisconsin ice invasion, and the underlying formations are exposed only in isolated localities. In this region the Lance formation has been mapped about 50 miles eastward from Bismark but is presumably unrecognizable farther northeast.[1] It seems a very reasonable supposition that the lower fresh-water part of the Lance may not extend very far to the east and that where it is absent marine sedimentation was continuous throughout the time when the fresh-water Lance was being deposited elsewhere.

Beyond the limits of Cannonball marine sedimentation on the west the Fort Union formation rests directly on the fresh-water beds of the Lance, and, except along the badlands of Little Missouri, no definite line of demarkation between the two has been drawn. In general the Lance is distinguished lithologically by a greater proportion of somber-colored shale and sandstone, the large number of the bog iron ore nodules and layers, by irregularity of deposition, and by a paucity of lignite. The Fort Union, on the other hand, contains a larger proportion of yellow and white sandstone, quartzite, and thin limestone, and almost everywhere contains a large number of lignite beds, some of which are very thick and persistent. The rocks of this formation are fine-grained and regularly bedded and give rise to angular topographic forms. In a large area in eastern Montana, mapped by the U.S. Geological Survey in 1910, it was found that the color change from somber below to yellow above does not take place everywhere at

[1] A. G. Leonard, "The Geological Map of North Dakota," *Quar. Jour. Univ. N.Dak.*, IV (1913), 4.

the same horizon. One example is cited by Calvert[1] where in 2 miles the somber beds transgress a distance of 200 feet upward into the yellow.

At present the only accepted criterion for the distinction of the Lance formation from the Fort Union in areas where the Lebo andesitic member of the Fort Union[2] and the Cannonball marine member of the Lance are absent is the presence in the former of Triceratops and its associated fauna. On the basis of its fauna and flora the Lance formation is approximately correlated with the Denver and Arapahoe formations of the Denver Basin, Colorado,[3] and with a part of the "Upper Laramie" described by Veatch[4] in Carbon County, Wyoming. In both these regions the supposed Lance equivalents usually rest unconformably upon a thick series of fresh- and brackish-water beds—the Laramie of the Denver Basin and the "Lower Laramie" of Carbon County, Wyoming. Thus the correlations made on the basis of the vertebrate fauna and the flora indicate that the Laramie is older than the Lance. In the Denver Basin the Laramie rests conformably on the Fox Hills sandstone.

A comparison of the stratigraphic section in the Denver Basin and adjoining regions with that in North and South Dakota presents certain anomalies which are not easy of interpretation. Knowlton[5] has maintained that there is an unconformity throughout the northeastern Plains region between the Fox Hills sandstone and the Lance which represents the time equivalent of the whole of the Laramie of the Denver Basin as well as an unconformity

[1] W. R. Calvert, "Geology of Certain Lignite Fields in Eastern Montana," *U.S. Geol. Survey Bull. 471*, p. 197, 1912.

[2] G. S. Rogers, "The Little Sheep Mountain Coal Field, Dawson, Custer, and Rosebud Counties, Montana," *U.S. Geol. Survey Bull. 531*, pp. 168–172, 1913.

[3] S. F. Emmons, Whitman Cross, and G. H. Eldridge, "Geology of the Denver Basin in Colorado," *U.S. Geol. Survey Mon. 27*, 1896.

[4] A. C. Veatch, "On the Origin and Definition of the Geologic Term 'Laramie,'" *Am. Jour. Sci.*, 4th ser., XXIV (1907), 18–22; *Jour. Geol.*, XV (1907), 526–49.

[5] F. H. Knowlton, "The Stratigraphic Relations and Paleontology of the 'Hell Creek Beds,' 'Ceratops Beds,' and Equivalents, and Their Reference to the Fort Union Formation," *Washington Acad. Sci.*, *Proc.*, XI (1909), 179–238; "Further Data on the Stratigraphic Position of the Lance Formation ('Ceratops Beds')," *Jour. Geol.*, XIX (1911), 358–76.

above the Laramie. Numerous localities in Wyoming, Montana, and North and South Dakota are described by Knowlton where this unconformity is supposed to be shown, (1) by erosion channels in the Fox Hills sandstone; (2) by discordance of dips between the two formations; or (3) by the great variation in thickness of the Fox Hills sandstone. In each case, however, the facts can be explained on a theory of practically continuous and unequal deposition.[1] Furthermore, the presence in the Dakotas of a marine fauna very similar to that of the Fox Hills sandstone, overlying the fresh-water sediments of the Lance, renders less tenable the theory of an unconformity of any importance at the base of the Lance, since the open sea must have persisted throughout Lance time in a region not very remote from western North Dakota.

Two other explanations have been suggested, the first that the fresh-water Lance strata of the Dakotas are the equivalents of the Laramie, Arapahoe, and Denver formations of the Denver Basin and of both the "Upper Laramie" and "Lower Laramie" of Carbon County, Wyoming. In this case the unconformity at the base of the Arapahoe is not of widespread general importance. The other and, to the mind of the writers, the more plausible hypothesis is that the Fox Hills sandstone of the Denver Basin is not the time equivalent of that in the type locality in South Dakota but is considerably older. The Fox Hills sandstone is, under this supposition, merely a near-shore phase of the sedimentation of the Montana sea. This sea persisted to much later time in the Dakotas than in Colorado, and the sedimentation was practically continuous until the close of the Fort Union. The possibility of finding a region where sedimentation was continuous throughout Laramie, Arapahoe, and Denver time was pointed out by Cross in 1909.[2] This deposition, it would seem, continued in the Dakotas in the open sea until near the close of Laramie and probably also throughout the period of erosion following the Laramie. Then the sea

[1] T. W. Stanton, "Fox Hills Sandstone and Lance Formation ('Ceratops Beds') in South Dakota and Eastern Wyoming," *Am. Jour. Sci.*, 4th ser., XXX (1910), 172–88.

[2] Whitman Cross, "The Laramie Formation and the Shoshone Group," *Washington Acad. Sci., Proc.*, XI (1909), 35.

withdrew still farther to the east and with the beginning of Lance time continental sedimentation was inaugurated over a large part of the eastern Rocky Mountain and Great Plains region. The last advance of the sea into the region of the Dakotas was approximately contemporaneous with the final dying-out of the Cretaceous dinosaur fauna, but there was no marked break in the continuity of sedimentation until the close of the Fort Union. Then the area suffered deformation and the erosion of thousands of feet of strata before the deposition of the later formations which rest unconformably on the Fort Union and all the underlying formations.

IS THE LANCE FORMATION TERTIARY OR CRETACEOUS?

In a general consideration of the age of the late Cretaceous or early Tertiary formations in the Rocky Mountain and Great Plains regions there are five lines of evidence which must be taken into consideration, and the decision at which any geologist will arrive will depend largely upon the relative importance which he gives to one or another of these lines of evidence. In a paper dealing primarily with the stratigraphy of a limited area in the Great Plains the larger and more general problem can be outlined only very briefly.

In the opinion of the writers the greatest weight should be given to the evidence of diastrophism. This, however, is of undetermined significance. The greatest break in the sedimentary record comes at the close of the Fort Union where there was everywhere, so far as known, in the Rocky Mountain and Great Plains regions an interval of long erosion and in many places of extensive folding. An earlier but apparently much less extensive break in the continuity of sedimentation is represented by erosional unconformities of greater or less extent at the base of the Raton, Dawson, Arapahoe, and "Upper Laramie" formations in New Mexico, Colorado, and Wyoming. An attempt to show that this unconformity is general throughout the northern plains region has, as previously pointed out, met with little success. In the Gulf and Atlantic Coastal Plains provinces the interval between the Cretaceous and Tertiary systems is marked by a great change in marine faunas and presumably by a profound break in the sedimentary

record. Obviously, if a definite correlation can be made of this unconformity in the Coastal Plain region with an unconformity in the Rocky Mountain region a long step has been taken toward settling the problem for all of the western region.

This leads directly to a second line of evidence—the fossil flora. The recent work of F. H. Knowlton[1] and E. W. Berry has shown a close relationship between the floras of the Midway and Wilcox formations of the Gulf region, which is indisputably of age, with the floras of the Raton and the Denver formations Eocene of New Mexico and Colorado. This evidence, together with the fact that the Raton and the Denver and Arapahoe formations rest unconformably on underlying formations which contain totally different fossil floras, has led Knowlton to the conclusion that the Raton and Denver formations are of Tertiary age. This conclusion is now accepted by the U.S. Geological Survey, but some geologists question the correlation, as it depends mainly on the evidence of the fossil flora.

When we turn to the evidence gathered from a study of the marine invertebrates we arrive at a diametrically opposite conclusion. As has been pointed out in preceding pages, the marine fauna of the Lance formation is distinctly a Cretaceous fauna very closely related to the fauna of the Fox Hills sandstone. It has been pointed out too that the Lance and Fort Union formations should be grouped together both on account of the close relationship of their fossil floras and because in places no lithologic separation of the two can be made. Thus the presence of a marine Cretaceous fauna in the Lance indicates that the Lance and probably also the Fort Union should be placed in the Cretaceous system. This would make the line between the Cretaceous and Tertiary correspond with the major unconformity above the Fort Union. It must be admitted that in the Rocky Mountain and the Great Plains regions no marine fauna of undoubted Tertiary age is known with which the Lance fauna may be compared, and there is a possibility that in this area the Fox Hills marine

[1] F. H. Knowlton, "Results of a Paleobotanical Study of the Coal-bearing Rocks of the Raton Mesa Region of Colorado and New Mexico," *Am. Jour. Sci.*, 4th ser., XXV (1913), 526–30.

fauna lived over from the Cretaceous into the Tertiary and thus became contemporaneous with the totally different marine faunas of the Gulf and Atlantic coasts, but if this did happen it is the first recorded case of such contemporaneous Cretaceous and Eocene faunas.

The evidence of the reptilian fossils in the Lance formation supports the evidence of the marine fauna. This—the Triceratops fauna—is the last of the great dinosaur faunas of the Mesozoic and for that reason has been considered generally by vertebrate paleontologists as sufficient evidence of the Mesozoic age of the formation. It is, however, just as probable that the Ceratopsian dinosaurs lived over into the Tertiary as that the marine Fox Hills fauna did so.

There remains yet to be considered the evidence of the mammalian fauna of the Fort Union formation. This fauna, of course, has its closest affinities with the succeeding Eocene faunas, since practically no mammalian fauna is known in the later Cretaceous formations. There is here the possibility that the Tertiary mammalian faunas began their development before the close of the Cretaceous. Lee[1] has recently described a find of a bone of a Creodont mammal associated with Ceratopsian bones in the Dawson arkose which shows that Ceratopsian dinosaurs and mammals of Tertiary aspect must have lived at the same time.

Such in outline is the Lance problem as it stands today. A recent detailed consideration of all of the evidence has led to a decision by the U.S. Geological Survey that the Denver and Arapahoe, Dawson, and Raton formations in Colorado and New Mexico all be placed in the Tertiary system. This decision was based primarily on the correlation of these formations with the Wilcox formation of the Gulf region on the evidence of their fossil floras and also on the consequent correlation of the unconformities in the two regions. Although the Lance formation is believed to be of about the same age as the Denver, Raton, and "Upper Laramie," it is classified by the U.S. Geological Survey

[1] W. T. Lee, "Recent Discovery of Dinosaurs in the Tertiary," *Am. Jour. Sci.*, 4th ser., XXXV (1913), 531–34.

as Tertiary(?), the doubt being thus expressed on account of the Cretaceous character of the Cannonball marine fauna. The writers believe that greater weight should be given to the evidence of the marine faunas and the dinosaurs, and that, in view of the strong evidence presented by these faunas, the correlations made on the basis of fossil floras should not be considered as conclusive.[1]

[1] This paper was submitted for publication in the *Journal of Geology* in June, 1914. Further studies on the Lance problem have strengthened the conviction of the writers that the Lance is Cretaceous.

VARIATIONS OF GLACIERS. XIX[1]

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The following is a summary of the *Eighteenth Annual Report* of the International Committee on Glaciers.[2]

THE REPORT OF GLACIERS FOR 1912

Swiss Alps.—The tendency of the Swiss glaciers to advance, which began in 1910, and which disappeared in 1911, following the wonderfully clear summer of that year, again became evident in 1912 (and more definitely than in 1910), probably on account of the very cold and wet summer of 1912. Although the number of glaciers retreating is still larger than the number of those advancing the former number is diminishing and the latter increasing.

Eastern Alps.—The heavy snowfall, the low summer temperature, and the hazy condition of the atmosphere reduced the melting during the summer to such an extent that even in August the glaciers were still covered with snow to their ends. One would expect, as a consequence, a marked advance of the glaciers, but this was by no means the case; of the thirty-four glaciers observed, fully twenty-eight were clearly in retreat, three were stationary and only three (and they were small) were advancing. Whether the advance of these three was due to the conditions holding during this one summer, or whether it is the forerunner of a general advance must be determined by the future. We can summarize by saying that the retreat of the glaciers of the Eastern Alps continues, but to a less marked degree than last year.

Italian Alps.—Observations were made in the Piedmont, the Lombard, and the Venetian Alps; they were greatly interfered with by the heavy snowfall, but the glaciers still continue to retreat.

French Alps and Pyrenees.—The tendency of the glaciers to advance, shown in 1911, has not been maintained in 1912. Some

[1] Earlier reports appeared in the *Journal of Geology*, III–XXI.

[2] *Zeitschrift für Gletscherkunde*, VIII (1913), 42–62.

glaciers show slight advances, others slight retreats. The snowfall was very heavy, but it has not yet affected the extent of the glaciers. In the basin of the Gave de Pau, Pyrenees, the snowfall in the winter of 1911–12 was small, but on account of the small summer melting very many of the glaciers remained covered with snow at the end of the summer. The glaciers increased in thickness especially in their reservoirs. On the whole, the glaciers observed have shown for several years a definite tendency to advance.

Swedish Alps.—The two glaciers observed show some slight advance.

Norwegian Alps.—All the glaciers of the Jotunheim are retreating. In the Folgefon and Jostedalsbrae, along the west coast, eighteen glaciers are retreating and six are advancing. In the more northerly regions eight are retreating and three are advancing. A larger proportion of the glaciers under observation were retreating in 1911 than in 1912.

Greenland.—A small tongue of the inland ice near Disco Bay experienced an advance which culminated in July, 1912. Later in the summer the ice had retreated. The large tongue in the same neighborhood, called the Ekip Sermia, seemed to be stationary. The number of icebergs in the Jacobshavn Fiord was becoming smaller. Three small glaciers on Disco Island, first described by Chamberlin in 1894, and mapped in 1897, have since that date retreated between 30 and 80 meters. There are also other indications of the diminution of the glaciations. Recent moraines in front of several glaciers in Sermilik Fiord show that these glaciers are retreating.

REPORT OF THE GLACIERS OF THE UNITED STATES FOR 1913

The Arapahoe Glacier, in Colorado, shows no appreciable change (Henderson).

Professor Lawrence Martin sends me the following information regarding the variations of Alaskan glaciers in 1913:

Glacier Bay.—Several glaciers in this fiord were studied by the writer in 1913, under the auspices of the National Geographic Society. Grand Pacific Glacier, which retreated 7,425 feet between June 1 to August 1, 1912, had advanced 4,000 feet by September 9, 1913, so that it again terminates south

of the International Boundary. De Margerie Glacier and the Reid and Lamplugh glaciers at the upper end of Glacier Bay were likewise active and advancing, the two latter moving forward half a mile from 1911 to 1913. Rendu Glacier retreated about 4,100 feet from 1911 to 1913. The adjacent cascading glacier, which was advancing in 1911, was retreating, and no longer reached tide water in 1913. The southwesternmost tributary of Rendu Glacier advanced in 1913 and pressed out the medial moraines of the Rendu. A cascading glacier on the east was probably also advancing.

Adams Glacier in Muir Inlet retreated about 3,575 feet from 1907 to 1913. Muir also continued to recede. It was more accurately mapped in 1913 than at any time since 1892. Its total recession between these dates was about $7\frac{3}{4}$ miles. Its decrease in thickness is notable; according to Reid's Survey, the ice surface was 1,500 feet above sea-level in 1892 at the site of the ice-front of 1913; and nunataks, 1,050 and 1,150 feet high, were, at the earlier date, covered by ice to depths of 400–600 feet. Soundings in Muir Inlet in 1913 revealed a depth of 1,128 feet at a point where the ice surface was 1,250 feet above sea-level in 1892. Thus the total thickness of Muir Glacier $5\frac{5}{8}$ miles from its 1892 terminus is now known to have been 2,378 feet. Soundings likewise show that in Tarr Inlet (at the upper end of Glacier Bay) the ice of Grand Pacific Glacier at a point 12 miles from the terminus of 1894 was at least 2,500 feet thick.

Near Lynn Canal.—R. G. McConnell reports that the Rainy Hollow or Sullivan Glacier had advanced about 2,000 feet before it was observed in 1910 by Webster Brown, making a total advance of about three-quarters of a mile in less than ten months. It receded slightly between 1910 and 1913. Jarvis Creek Glacier likewise advanced in 1910. The detailed map of the Mendenhall, Eagle, Herbert, and adjacent glaciers east of Lynn Canal,[1] is accompanied by cursory notes on these ice tongues by Knopf.

Yakutat Bay.—Since 1910 Hidden Glacier has retreated 400–500 feet at the terminus and thinned 150 feet or more at the north margin. The tidewater terminus of Nunatak Glacier retreated slightly from 1912 to 1913. The adjacent cascading glacier advanced several hundred feet from 1910 to 1913. Hubbard Glacier receded slightly. Turner Glacier suffered a more pronounced diminution, especially at its south margin.

Mount St. Elias Region.—North of Mount St. Elias the Logan-Chitina Glacier changed from spasmodic, earthquake-stimulated activity in 1912 to stagnation in 1913, as reported by D. W. Eaton and J. D. Craig of the International Boundary Surveys. Its margin receded 30–50 feet during the year. A number of the ice tongues south of Logan Glacier were first mapped in 1913. The surveyors saw no unusual activity between Logan Glacier and Mount St. Elias but report that ablation during cloudy, rainy weather in July lowered the ice surface at a measured rate of between 3 and 4 inches per day.

[1] *Bull. No. 502, U.S. Geol. Survey,* 1912.

The new Icy Bay at the western border of Malaspina Glacier was mapped in 1913 by A. G. Maddren of the United States Geological Survey. The recession of Guyot Lobe now amounts to nine or ten miles. Maddren's map indicates that the Bering Glacier, which has hitherto been considered a single piedmont ice mass, is really two separate piedmont bulbs (Coast Survey Chart 8002, edition of March 12, 1912). A similar condition at Grand Plateau Glacier was noted from close inshore by the special steamer of the International Geological Congress of 1913. In both cases the earlier mapping was in error, for there has been no recession of either the Grand Plateau Glacier or the Bering Glacier in historical times.

Copper River Canyon.—The Childs Glacier has retreated slightly and its northern margin, as measured by Caleb Corser, is now 1,500 feet from the railway bridge on the Copper River & Northwestern Railway. In spite of an advance of more than 60 feet in the interval, its net change, from June 16, 1911, to July 12, 1913, was a recession of 26 feet, and from the latter date to November 7, 1913, it retreated between 100 and 150 feet more. Allen Glacier, after a slight advance in 1912, began to recede in 1913. Between August 26, 1910, and July 12, 1913, it advanced 2650 feet, or four-tenths of the distance to the railway north of it. From the latter date to October 28, 1913, Corser's measurements show that its northwestern border retreated 300–700 feet. Schwan Glacier is said by a prospector to have advanced in 1912.

Wrangell Mountains.—In connection with her successful ascent of Mount Blackburn in 1912,[1] Miss Dora Keen made important observations of snowfall and temperature, giving an idea of the precipitation that supports the large Kennicott Glacier. The total snowfall for thirty-three days (April 22 to May 24) was 40 feet (packed to 20 feet). During this period the temperature at altitudes of from 2,000 to 16,140 feet above sea-level ranged from $-6°$ F. to $+70°$ F. At her base camp, altitude 5,500 feet, three feet of snow fell in thirteen consecutive days, while at a camp 6,900 feet higher the snowfall during the same period, which included four days of hot sun, was very much greater. It was measured by the burial of bamboo poles 12 feet long in front of a shelter excavated in the snow and may be estimated at 30 feet (packed to 10 feet). At the summit of Mount Blackburn the snowfall seems to be less, though it may be all carried away by the winds. An important part of the alimentation of Kennicott Glacier comes in great snow avalanches from the steep walls of its valley. Some of these avalanches, which are most abundant between noon and 4:00 P.M., advance three miles from the valley wall, or the whole width of the glacier.

Prince William Sound.—Barry Glacier in Harriman Fiord retreated an amount varying from about 2,500 feet on the western side to at least 7,200 feet on the eastern border between Martin's visit in July, 1910, and Johnson's in 1913. Coxe Glacier, formerly a tributary of the Barry is now independent.

[1] *World's Work*, XXVII (1913), 80–101.

Cook Inlet.—A glacier at the head of Tuxedni or Snug Harbor seems to have advanced between 1904 and 1911. Coast Survey Chart 8554 shows some of the glaciers on Mount Iliamna for the first time and new details of the glaciers on Mount Douglas.

In 1913 Archdeacon Stuck found that a marked change had taken place in one of the great ridges of Mount McKinley since Parker's ascent in 1912; the snow and ice had broken away from a larger portion of the ridge, owing, possibly, to the earthquake of July 6, 1912.

In the absence of M. Charles Rabot, the retiring president of the Commission International des Glaciers, his report to the Twelfth International Congress of Geologists at Toronto was presented by Mr. Emile de Margerie.[1] It may be summarized as follows:

At present the glaciers throughout the world are in general retreating. Though in some regions, as Iceland, the retreat is slight, in others, such as Alaska, Norway, and the Alps, it is very marked. Thus in the Pelvoux *massif*, several small glaciers have entirely disappeared in the last thirty years.

The retreat, which has lasted for a century in Norway, and for fifty or sixty years in the Alps has been interrupted by small temporary advances. In Norway an advance in the last few years is dying out; in the French Alps and in the Pyrenees a secondary advance is developing.

A careful search of old documents has revealed the variations of the Chamonix glaciers since 1580. A strong advance occurred during the last years of the sixteenth century; others in 1643, in 1663, and in 1716; the last culminated only about 1741. The variations since then were already known: a new advance began near the end of the eighteenth century, reached its maximum in 1818–20, and continued, but in a mild form, until the middle of the nineteenth century; since then the glaciers have been in retreat. It seems that the advance about 1600 was greater than had occurred for several generations before that date; and that, since then, the glaciers have not been as reduced as they are at present. In Norway, also, branches of the Jostedalsbrae and of the Svartisen advanced strongly in the first quarter of the eighteenth century; and in Iceland an increase of the glaciation seems to have existed

[1] The report is printed in the *Comptes rendus* of the Congress, pp. 144–48, and in *Zeitschrift für Gletscherkunde*, VIII (1914), 263–69.

since the tenth century. These records lead to the conclusion that glaciers are subject to variations of two orders: primary variations of great amplitude and of unknown duration; and, superposed on them, secondary variations, which, in the Alps, appear to follow Brückner's law. These oscillations are due to climatic variations, and it is advisable that brief reports of the snowfall should accompany the records of glacier variations.

Since the last congress the commission has suffered severely by the deaths of F. A. Forel, who may be looked upon as the originator of the study of glacier variations; of K. J. V. Steenstrup, who has done such excellent work in Greenland; and of R. S. Tarr, well known for his studies in Alaska. M. Paul L. Mercanton, of Lausanne, succeeds M. Forel as the ordinary member for Switzerland; M. Paul Harder takes the place of M. Steenstrup representing Denmark; and Mr. Alan G. Ogilvie now represents Great Britain in the place of Mr. Freshfield, who has retired and become a corresponding member. Other corresponding members elected recently are: Messrs. J. Rekstad, Adolf Holl, A. O. Wheeler, Lawrence Martin, and F. E. Matthes.

The officers of the commission elected to serve until the next meeting of the International Congress of Geologists are: Honorary President, Prince Roland Bonaparte; Active President, Dr. Axel Hamberg; Secretary, Dr. Paul L. Mercanton.

A NEW GENUS AND SPECIES OF AMERICAN THEROMORPHA

MYCTEROSAURUS LONGICEPS

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The past summer, Mr. Herman Douthitt, of the University of Chicago paleontological expedition, discovered on Mitchell Creek, Texas, in the horizon that has yielded various specimens of *Pantylus* and the type specimens of *Broiliellus* and *Glaucosaurus*, a broken skull and parts of the skeleton of a small reptile which, at the time, were thought to belong to the genus *Varanops*. Although it resembles that genus in shape and general characters, a more careful examination disclosed a new and peculiar form of Theromorpha.

The specimen is of a nearly white color, inclosed in a rather hard, siliceous red clay, from which the bone readily separates, leaving impressions as though made in wax. The skull had suffered a little from compression, but is otherwise undistorted. It was originally quite complete, but, as found, some of the bones had separated from the matrix and been lost; and the tip of the premaxillæ was gone. This partial loss of the bony structure, on one side or the other, while detracting from the appearance of the specimen, has made vividly apparent nearly every suture; others are clearly shown in the bone itself. The sutures of the inner side of the cranial bones are not always quite the same as on the outside; as an instance of which, the shape and size of the postfrontal of *Pantylus* is quite as Case and Huene figured it, while on the outside Broom gave the form correctly; and the quadratojugal of the same genus is nearly twice as broad on the inside as on the outside. In the figures given herewith (Figs. 1, 2) I have widened the face a little in top view, perhaps not quite enough. I have depended in no instance on any line indicating a suture unless it is precisely corroborated on the two sides, precluding the danger of mistaking cracks for

sutures. These figures will, I trust, render a detailed description for the most part unnecessary. The points of interest meriting

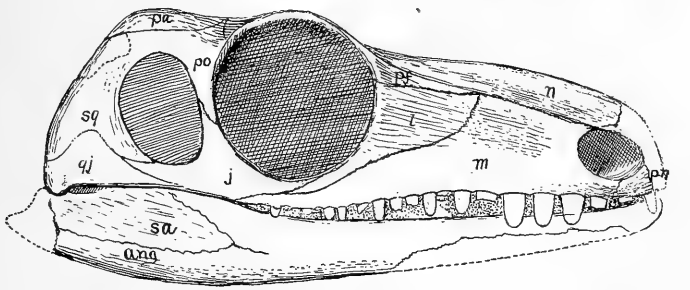

FIG. 1.—*Mycterosaurus longiceps* Will. Skull from the side. *n*, nasal; *pm*, premaxilla; *m*, maxilla; *l*, lacrimal; *pf*, prefrontal; *pof*, postfrontal; *po*, postorbital; *j*, jugal; *qj*, quadratojugal; *sq*, squamosal; *pa*, parietal; *f*, frontal; *ds*, dermosupraoccipital; *so*, supraoccipital; *sa*, surangular; *ang*, angular.

discussion are the nares, orbits, lacrimals, the boundaries of the temporal opening, and the structure of the occiput.

The nares are unusually large, as are also the orbits. The latter are nearly circular in outline, and look almost directly outward. They project above the frontal region, which is concave transversely between them. Within the orbits the broad smooth upper surface of the palatines is apparent, as also the narrow pterygoid; both bear small, conical teeth. The lacrimals do not reach the nares, as has been assumed without positive proof for some of the American Theromorpha. On the two sides not only do the sutural lines agree exactly, but the smooth, clean surface in front shows no trace of a suture on either side.

The jugo-postorbital arch is slender; on the right side it is bent inward, separating at the suture; on the left side it is broken in two or three places. The quadratojugal is definitely shown on each side. Because of the squamous

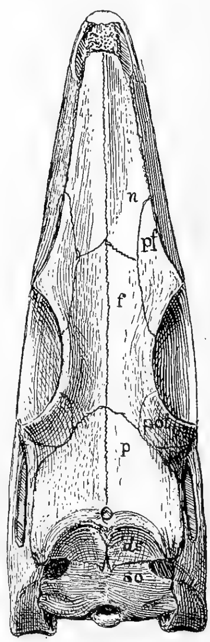

FIG. 2.—*Mycterosaurus longiceps*. Skull, from above. Explanations as in Fig. 1. Both figures natural size.

overlap of the squamosal one cannot be quite sure of the course of the suture externally, but it descends no lower than I have figured it. Above, the squamoso-postorbital arch is shown very decisively on each side, but I cannot be sure of the suture separating the two bones, and have, therefore, omitted it. I find no definite evidence of a tabulare or supra-temporal in this region, though it is not at all improbable that one or both are present. Posteriorly the squamosal covers the quadrate broadly. The temporal opening is moderately large and oval in form. It is bounded, it will be seen, almost precisely as in *Sphenodon*. The parietal on each side is slightly separated from the squamoso-postorbital arch, indicating one way in which the upper temporal opening in *Sphenodon* may have arisen. The parietal foramen is situated almost at the extreme posterior end of the parietals, and very close to the dermosupraoccipitals. On each side the parietals are curved downward to near the top of the quadrate, but I find no evidence of a suture separating the extremity of the processes.

The occipital surface is but little injured and quite clean of matrix. On the upper part of the steeper declivity, between the descending processes of the parietals, there is a thin, concave bone on each side, lying against the back part of the parietals, and somewhat separated in the middle line above by a descending process of the same bones. Below, each bone diverges a little to fit into a small groove on the upper external surface of the supraoccipital. The suture between these bones and the supraoccipital on each side is very conspicuous. The bones extend outward to the extremity of the posterior parietal processes. There is a small but distinct post-temporal opening on each side between the supraoccipital, dermosupraoccipital, and squamosal.

On the large surface below the dermosupraoccipitals and the temporal fossae I can distinguish no sutures. In shape, the surface is very much like that of *Dimetrodon*, as shown in the accompanying figures (Figs. 3, 4) made from a viewpoint at right angles to the plane of the supraoccipital. The paroccipital process is separated from the upper, supraoccipital process in the same way; and their union with the quadrate is very similar in both genera. I am aware that this bone in *Dimetrodon* has been differentiated into the dermo-

supraoccipital and supraoccipital, in addition of course to the paroccipitals, exoccipitals, and petrosal, but I can find no evidence whatever for the inclusion of the first-mentioned bone in the complex. I have sectioned various specimens, and studied numerous others. If, then, these bones correspond—and they surely do—there are no dermosupraoccipitals in *Dimetrodon*, or they have been reduced to the merest vestige, as in the modern gavials. As will be seen by the figure (Fig. 4, *po*) the surface of the supraoccipital articulating with the parietals above is cartilaginous, corresponding to the condition found in the dinosaurs, crocodiles, lizards, etc. The exoccipitals in *Mycterosaurus* surround the foramen magnum and are loosely attached, as in *Dimetrodon*. From below, not much is visible. The descending process of the paroccipital for the stapes is visible from behind, but I do not discover either the stapes or the tympanic, both of which are present in *Dimetrodon* (Fig. 4, *t*, *st*). This last figure is introduced here anticipatory of a full discussion of the cranial anatomy of this and other genera, which will be published later.

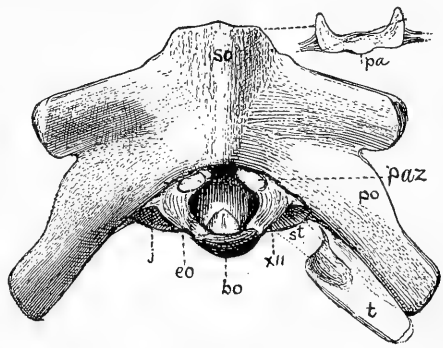

FIG. 3.—*Dimetrodon*. Occipital complex, obliquely from behind. *so*, supra-occipital; *paz*, proatlantal zygapophysis; *po*, paroccipital; *st*, stapes; *t*, tabulare; *bo*, basioccipital condyle; *eo*, exoccipital; half natural size.

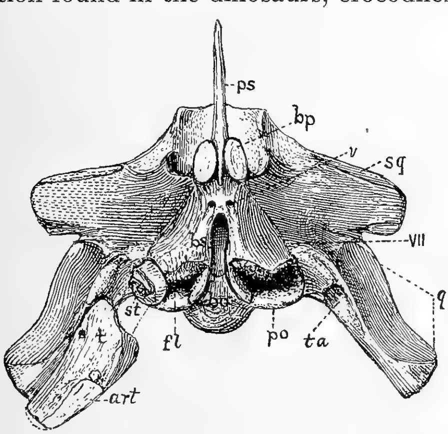

FIG. 4.—*Dimetrodon*. Occipital complex, view opposite to that of Fig. 1. *ps*, parasphenoid; *bp*, basipterygoid process; *bs*, basisphenoid; *bo*, basioccipital; *st*, head of stapes; *t*, tympanic; *ta*, articular pit in paroccipital for tympanic; *art*, articular surface; *po*, paroccipital.

The teeth in *Mycterosaurus* are characteristically different from those of *Varanops*, *Ophicaodon*, *Dimetrodon*, or *Edaphosaurus*. Altogether there are about eighteen, possibly one more, on each maxilla, the first four or five the largest. They are moderately elongated and flattened, with a rather obtuse apex.

Of the skeleton of this specimen, there are fragmentary vertebrae, and a coraco-scapula. The posterior coracoid is well ossified and not distinguishable from the anterior. It has a straight line below and is rather narrow. The interclavicle appears to be of the usual form, and very unlike the extraordinary interclavicle of *Pantylus*. The vertebrae, so far as they are preserved and prepared, are like those of *Varanops*. The spine is thin, and not more than two or three times the height of the centra.

Relationships.—There can scarcely be dissent from the opinion that *Mycterosaurus* is related to *Dimetrodon*, and should find its natural place in the same suborder, the Pelycosauria, though in a different family. The skull and vertebrae are more primitive, but both have the same relations of the bones, the same temporal opening, and the same elongated, narrow skull. The lower jaw lacks the inflected angular process below, but that is a specialization that would hardly be expected in the more primitive form. I am aware that this peculiarity has been largely relied upon as indicative of the relationships between the American and African Theromorpha, but I am skeptical of its value. No other American genus shows it, except, in a lesser degree, *Sphenacodon* and *Edaphosaurus*, and these are all highly specialized animals with elongated spines.

The family determination of *Mycterosaurus*, until more of the skeleton is known, is doubtful; it may provisionally be placed in the Poliosauridae. Of the American reptiles, nearly every specimen that can be differentiated by decisive characters represents a distinct genus; and the majority of genera, when fully known, are more or less justly placed in distinct families. The following genera, it seems to me, are rightly separated in different families; a fuller knowledge may require further division.

Sphenacodontidae
- Clepsydrops Cope
- Sphenacodon Marsh
- Dimetrodon Cope

Ophiacodontidae
  Ophiacodon Marsh
  Theropleura Cope
Poliosauridae
  Poliosaurus Case
  Varanosaurus Broili
  Varanops Williston
  ?Mycterosaurus Williston
  ?Arribasaurus Williston
Edaphosauridae
  Edaphosaurus Cope
Caseidae
  Casea Williston

Together with *Trichasaurus* Williston, *Scoliomus* Williston and Case, *Poecilospondylus* Case, *Glaucosaurus* Williston, *Elcobresaurus* Case, of more doubtful position, these are all that can be definitely located among the American Theromorpha, that is zygocrotaphous reptiles, with the temporal vacuity below the squamoso-postorbital arch. *Ostodolepis* Williston is a synonym of *Pantylus* Cope.

USE OF THE SLIDE RULE IN THE COMPUTATION OF ROCK ANALYSES

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Certain phases of petrographic and metamorphic work require the conversion of rock analyses into the corresponding approximate mineralogical compositions. By ordinary methods of calculation this may be rather prolonged and call for a considerable expenditure of time and energy. Exact conversion is, of course, impossible, especially without any preliminary data as to the minerals actually present, and, in the case of such variables as the biotites, amphiboles, and pyroxenes[1] their chemical composition itself should be determined also. Nevertheless the results attainable are of sufficient value to justify the procedure.

Various methods have been suggested, each possessing more or less merit and marking some advance in this direction. The use of molecular ratios has been very general and has been described by Kemp,[2] Cross,[3] and Osann.[4] Harker[5] calculated decimal tables for the different minerals which very materially assist in converting oxide values into the corresponding mineral values, but which are not adapted to the reverse process which in some cases is of equal importance. Mead[6] devised a set of graphic tables, one for each mineral, by means of which the value of any radicle in a mineral might be converted into the corresponding value of any other radicle

[1] Cross, Iddings, Pirsson, and Washington, *Quantitative Classification of Igneous Rocks*, 1903, p. 145.

[2] J. F. Kemp, "The Recalculation of the Chemical Analyses of Rocks," *School of Mines Quarterly*, XXII (1900–1901), 75.

[3] Cross, Iddings, *et. al.*, *op. cit.*, Part III, "Methods of Calculation."

[4] A. Osann, *Beiträge zur chemischen Petrographie*, I. Teil. "Molekular-Quotienten zur Berechnung von Gesteinsanalysen," Stuttgart, 1903.

[5] A. Harker, private publication, Cambridge University Press, 1910.

[6] W. J. Mead, "Some Geological Short-Cuts," *Economic Geology*, VII, No. 2 (March, 1912), 136.

in that same mineral or to the total mineral value. Conversely, the value of any radicle could be determined from the total mineral value. This marked a decided step in advance, but as a complete table is necessary for each mineral the disadvantages are obvious.

Later in the same article[1] Mead describes a "geologist's slide rule," which is a much more efficient method of conversion, and which has certainly merited a wider consideration than it has apparently received. He has devised a rule for ordinary rock minerals and another one for ore minerals, comprising a total of nearly 100 minerals, whereby very rapid conversions and comparisons are possible, either from constituent to mineral or the reverse. The rules are of white celluloid, about ten inches in diameter and are exceedingly simple to manipulate. In using these rules certain features became apparent to the writer: (1) a rule is good only for the minerals printed thereon; (2) ordinary slide-rule computations are impossible on the logarithmic scale of the rule; (3) the rule is too large for ordinary pocket use and hence of limited value in the field. These statements are not intended to detract in any way from the value of the "geologist's slide rule," but led the writer to consider a way to broaden this excellent application of the logarithmic scale to geologic work.

In a recent publication Lindgren[2] gives a summary of the various methods of recalculation. The scheme most favored "consists of a co-ordinate system in which the abscissas represent the distance from the vein which may be taken as the origin of the acting solutions, and the ordinates represent the molecular ratios multiplied by 100, except for silica, for which the scale must be reduced to bring the diagram within convenient compass." This method is excellent for just such cases as cited,[3] but a diagram so constructed is limited to one definite zone or locality and does not have the range of broad comparisons and generalizations inherent in the straight line diagram.

In the conversions and comparisons of chemical analyses referred to, considerable multiplication and division are necessary

[1] *Ibid.*, p. 139.

[2] Waldemar Lindgren, *Mineral Deposits*, 1913. Chap. xxx.

[3] F. L. Ransome, *U.S. Geol. Survey Professional Paper 75*.

in addition to that obviated by the "geologist's slide rule." For this work slide-rule accuracy is sufficient. In extended studies of this nature the use of the rule may save from 50 to 90 per cent of the time required for such calculations. Hence, if these conversions could be simply accomplished on an ordinary slide rule, considerable time and energy would be saved and the rule might then attain the same merited popularity with geologists that it has already won in engineering lines. At the same time its compact nature lends itself more readily to field use.

For these reasons the writer submits the accompanying tables (Tables I and II) of mineral percentage compositions, with a method of use applicable to any ordinary slide rule, which he trusts may meet the foregoing conditions successfully. In addition, conversion into molecular values by means of tables such as given by Osann[1] can be so much more quickly accomplished on a slide rule than by ordinary arithmetical processes.

	I*	II†
SiO_2	76.87	48.45
TiO_2	0.07	0.53
Al_2O_3	13.10	18.47
Fe_2O_3	1.05	4.81
FeO		3.96
MnO	0.01	0.06
MgO	0.06	3.50
CaO	0.45	7.38
Na_2O	3.55	4.58
K_2O	4.83	7.78
P_2O_5	0.01	0.48

* R. A. Daly, *Igneous Rocks and Their Origin*, 1914, p. 34, No. 97 (alaskite).
† *Ibid.*, p. 32, No. 90 (leucitite).

These tables contain two sets of minerals, one the more common of the rock-forming variety, the other the more important ore minerals. Precise mineral composition is, of course, only theoretically attained, but the formulae given are taken from standard texts[2] and represent generally accepted values. An

[1] A. Osann, *Beiträge zur Chemischen Petrographie*, I. Teil, "Molekular-Quotienten zur Berechnung von Gesteinsanalysen," Stuttgart, 1913.

[2] E. S. Dana, *Textbook of Mineralogy;* Moses and Parsons, *Mineralogy, Crystallography and Blowpipe Analysis;* F. W. Clarke, "Data of Geochemistry," *U.S. Geological Survey Bulletin 491;* Albert Johannsen, *Determination of Rock-forming Minerals in Thin Sections.*

average value for specific gravity is given instead of the range usually stated, as this average is accurate enough for such computations, and in fact is more accurate than the reconstructed analyses. This is due to the fact that the rocks of unlike mineral composition may have identical chemical composition, and also due to the range in chemical composition of such minerals as the biotites, amphiboles, pyroxenes, garnets, spinels, etc. The use of the tables may be explained most easily by taking two rock analyses and performing two conversions such as commonly confront the petrographer.

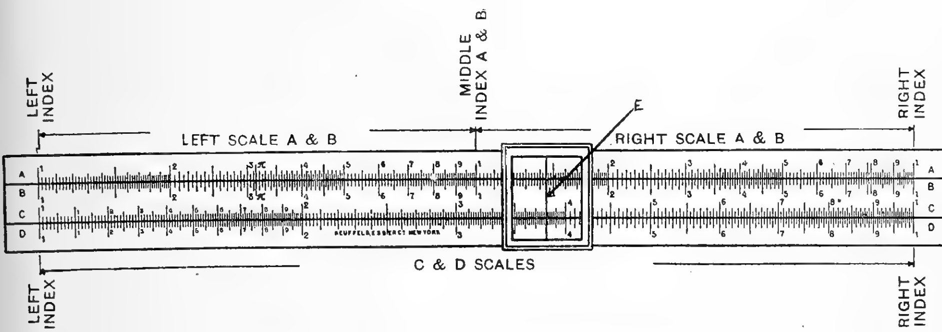

FIG. 1

The ordinary slide rule (see Fig. 1) has four graduated logarithmic scales, two on the immovable portion and two on the slide. For convenience these are referred to as "A," "B," "C," and "D," beginning with the upper one and reading downward. For brevity's sake the vertical line on the glass slide will be termed "E." A and B are duplicates, as are C and D. In using the mineralogical tables the scales C and D only are needed. First we shall proceed to reconstruct albite in Analysis I. The chemical analysis shows the presence of 3.55 per cent Na_2O. In Table I the percentage composition of albite is found to be

Na_2O—11.8; Al_2O_3—19.5; SiO_2—68.8

Set "E" on 118 on "D" and move "C" until 355 on this scale is under "E." Then if the corresponding amount of albite is desired.

move "E" to the left index on "D." The reading on "C" under "E" is 301, which means that 30.1 per cent of albite is equivalent to 3.55 per cent of Na_2O. Similarly, to find the amount of Al_2O_3 needed to combine with 3.55 per cent Na_2O in the albite molecule move "E" to 195 on "D." This will then cross "C" at 587, which indicates a requirement of 5.87 per cent of Al_2O_3. The sliding portion of the rule (carrying scales "B" and "C") has been set but once for both these readings. Now suppose that the SiO_2 value in albite corresponding to 3.55 per cent Na_2O is desired. In the table the SiO_2 percentage is 68.8. This value on "D" is not under "C" at all in the present setting. The next operation corresponds merely to doubling the length of the scale on "C." Place "E" on the right index of "C" and then move "C" to the right until the left index of "C" is under "E." Then move "E" to 688 on "D" and it is found to cross "C" at 20.7 which is the SiO_2 value required. A check on the correctness of these values thus obtained is to total the constituents and compare with the total obtained on "C."

Total on "C"	Separate Percentages
Albite—30.1 per cent	Na_2O—3.55 per cent
	Al_2O_3—5.87 " "
	SiO_3—20.7 " "
	Total—30.12 per cent

Now suppose we consider the potash in Analysis II which probably was present as leucite (7.78 per cent). Table I shows the composition of leucite as follows:

K_2O—21.5; Al_2O_3—23.3; SiO_2—55.2

Set "E" on 215 on "D" and move the slide so that 778 on "C" is under "E," or is opposite 215 on "D." Move "E" to the left index on "D" and it is found to cross "C" at 36.19, which is the percentage of leucite present. In order to get the corresponding value of Al_2O_3 move "E" to 233 on "D." In this position it crosses "C" at 8.43, the percentage of Al_2O_3. The silica value, 55.2 on "D," again falls under the blank space in the rule, so set "E" on the right index of "C," move the slide to the right so that the left

index of "C" is under "E" and then set "E" on 552 on "D." In this position it crosses "C" at 19.98.

Total on "C"	Separate Percentages
Leucite—36.19 per cent	K_2O—7.78 per cent
	Al_2O_3—8.43 " "
	SiO_2—19.98 " "
	Total—36.19 per cent

The reverse operation, or conversion of minerals to their constituent oxides, will now be described. Suppose we take the case just cited. A rock contains 36.19 per cent leucite. What are the corresponding values of K_2O, Al_2O_3, and SiO_2? The percentage composition of leucite is

K_2O—21.5; Al_2O_3—23.3; SiO_2—55.2

Set "E" over an index on "D," and since there is a choice of two such positions (one at each end of the scale) suppose the right index of "D" is chosen. Then move the slide so that the leucite percentage, 36.19 on "C," is under "E." It will be noted that of the three oxide values—21.5, 23.3, 55.2—only that of silica can be found on "D" under the slide in its present position. Hence set "E" on 552 on "D" and it will be found to cross "C" at 19.98, the percentage value of SiO_2. Then shift the slide so that 36.19 on "C" is opposite the left index on "D." Set "E" on 215 on "D" and it will cross "C" at 7.78, which is the potash value in the leucite percentage. Move "E" to 233 on "D" and it will cross "C" at 8.43, which is the Al_2O_3 value desired.

This explanation covers any ordinary case, including both the synthetical and the analytical processes. If the following point is borne in mind there should be no confusion. In the method as outlined, all percentage values from the table were read on "D," whereas all analytical values and their equivalents were read on "C." The opposite way is equally feasible, but if one way is always followed there is little excuse for confusion. Similarly, all these operations can be performed on scales "A" and "B," but the divisions are smaller and not so easily read.

TABLE I

Rock Minerals

Actinolite: 3.1—$CaMg_2FeSi_4O_{12}$: CaO—12.5; MgO—17.9; FeO—16.0; SiO_2—53.6
Aegirite: 3.55—$NaFeSi_2O_6$: Na_2O—13.4; Fe_2O_3—34.5; SiO_2—52.1
Albite: 2.63—$NaAlSi_3O_8$: Na_2O—11.8; Al_2O_3—19.4; SiO_2—68.8
Almandite: 4.05—$Fe_2Al_2Si_3O_{12}$: FeO—43.2; Al_2O_3—20.5; SiO_2—36.3
Alunite: 2.66—$KAl_3O_6H_9S_2O_8$: K_2O—11.4; Al_2O_3—37.0; SO_3—38.6; H_2O—13.0
Analcite: 2.25—$NaAlSi_2O_6H_2O$: Na_2O—14.0; Al_2O_3—23.2; SiO_2—54.6; H_2O—8.2
Andalusite: 3.18—Al_2SiO_5: Al_2O_3—62.9; SiO_2—37.1
Andesine: 2.68—Ab_3An_2: Na_2O—6.9; CaO—8.3; Al_2O_3—26.6; SiO_2—58.2
Andradite: 3.85—$Ca_3Fe_2Si_3O_{12}$: CaO—33.0; Fe_2O_3—31.5; SiO_2—35.5
Anhydrite: 2.94—$CaSO_4$: CaO—41.2; SO_3—58.8
Anorthite: 2.75—$CaAl_2Si_2O_8$: CaO—20.1; Al_2O_3—36.7; SiO_2—43.2
Apatite: 3.20—$Ca_4(CaF)P_3O_{12}$: CaO—55.5; P_2O_5—42.3; F—3.8
Augite: 3.31: CaO—20.9; MgO—12.6; FeO—6.9; Fe_2O_3—4.6; Al_2O_3—7.4; SiO_2—47.6
Barite: 4.45—$BaSO_4$: BaO—65.7; SO_3—34.3
Bauxite: 2.48—$Al_2O_3.2H_2O$: Al_2O_3—73.9; H_2O—26.1
Biotite: 2.95: K_2O—11.2; (Mg, Fe)O—19.2; Al_2O_3—24.4; SiO_2—43.1; H_2O—2.1
Calcite: 2.72—$CaCO_3$: CaO—56; CO_2—44
Chlorite (Penninite): 2.72: MgO—19.5; FeO—23.2; Al_2O_3—16.5; SiO_2—29.2; H_2O—11.6
Diopside: 3.33—$CaMgSi_2O_6$: CaO—25.8; MgO—18.6; SiO_2—55.6
Dolomite: 2.84—$CaMgC_2O_6$: CaO—30.4; MgO—21.7; CO_2—47.9
Enstatite: 3.17—$MgSiO_3$: MgO—40; SiO_2—60
Epidote (Pistacite): 3.38—$Ca_2(Al, Fe)_3HSi_3O_{13}$: CaO—23.5; Al_2O_3—24.1; Fe_2O_3—12.6; SiO_2—37.9; H_2O—1.9
Fayalite: 4.14—Fe_2SiO_4: FeO—70.6; SiO_2—29.4
Fluorite: 3.13—CaF_2: Ca—51.1; F—48.9
Forsterite: 3.25—Mg_2SiO_4: MgO—57.1; SiO_2—42.9
Gibbsite: 2.36—$Al(OH)_3$: Al_2O_3—65.4; H_2O—34.6
Grossularite: 3.57—$Ca_3Al_2Si_3O_{12}$: CaO—37.3; Al_2O_3—22.7; SiO_2—40.0
Gypsum: 2.32—$CaSO_4.2H_2O$: CaO—32.6; SO_3—46.6; H_2O—20.9
Halloysite: 2.1—$Al_2Si_2O_7.3H_2O$: Al_2O_3—36.9; SiO_2—43.5; H_2O—19.6
Hedenbergite: 3.6—$CaFeSi_2O_6$: CaO—22.2; FeO—29.4; SiO_2—48.4
Hornblende: 3.24: CaO—5.8; MgO—8.3; FeO—22.1; Fe_2O_3—16.3; Al_2O_3—10.5; SiO_2—37.0
Hypersthene: 3.45—$(Mg, Fe)SiO_3$: MgO—17.3; FeO—30.9; SiO_2—51.8
Kaolinite: 2.62—$Al_2Si_2O_7.2H_2O$: Al_2O_3—39.5; SiO_2—46.5; H_2O—14.0
Labradorite: 2.71—Ab_2An_3: Na_2O—4.6; CaO—12.3; Al_2O_3—30.0; SiO_2—53.1
Leucite: 2.48—$KAlSi_2O_6$: K_2O—21.5; Al_2O_3—23.3; SiO_2—55.2
Magnesite: 3.04—$MgCO_3$: MgO—47.6; CO_2—52.4
Melilite: 3.0: Na_2O—4.3; CaO—31.3; MgO—8.4; Fe_2O_3—11.2; Al_2O_3—7.1; SiO_2—37.7
Muscovite: 2.87—$H_2KAl_3Si_3O_{12}$: K_2O—11.8; Al_2O_3—38.5; SiO_2—45.2; H_2O—4.5
Nephelite: 2.6—$K_2Na_6Al_8Si_9O_{34}$: K_2O—7.7; Na_2O—15.1; Al_2O_3—33.2; SiO_2—44.0
Oligoclase: 2.66—Ab_3An_1: Na_2O—8.8; CaO—5.2; Al_2O_3—23.9; SiO_2—62.1
Orthoclase: 2.56—$KAlSi_3O_8$: K_2O—16.9; Al_2O_3—18.4; SiO_2—64.7
Serpentine: 2.56—$H_4Mg_3Si_2O_9$: MgO—43.0; SiO_2—44.1; H_2O—12.9
Staurolite: 3.70—$HFeAl_5Si_2O_{13}$: FeO—15.8; Al_2O_3—55.9; SiO_2—26.3; H_2O—2.0
Talc: 2.73—$H_2Mg_3Si_4O_{12}$: MgO—31.7; SiO_2—63.5; H_2O—4.8
Titanite: 3.52—$CaTiSiO_5$: CaO—28.6; TiO_2—40.8; SiO_2—51.7
Wollastonite: 2.85—$CaSiO_3$: CaO—48.3; SiO_2—51.7
Zircon: 4.69—$ZrSiO_4$: ZrO_2—67.2; SiO_2—32.8
Zoisite: 3.31—$HCa_2Al_3Si_3O_{13}$: CaO—24.6; Al_2O_3—33.7; SiO_2—39.7; H_2O—2.0

Oxides
Al_2O_3: Al—53.03; O—46.97
BaO: Ba—89.58; O—10.42
CaO: Ca—71.47; O—28.53
CO_2: C—27.27; O—72.73
Cr_2O_3: Cr—68.42; O—31.58
CuO: Cu—79.51; O—20.49
FeO: Fe—76.34; O—23.66
K_2O: K—83.02; O—16.98
MgO: Mg—60.32; O—39.68
MnO: Mn—77.44; O—22.56
Na_2O: Na—74.19; O—25.81
PbO: Pb—92.83; O—7.17
P_2O_5: P—43.69; O—56.31
SO_3: S—40.05; O—59.95

TABLE II

Ore Minerals

Iron	Arsenopyrite: 6.05—$FeAsS_2$: Fe—34.3; As—46.0; S—19.7
	Chromite: 4.44—$FeCr_2O_4$: FeO—32; Cr_2O_3—68
	Goethite: 4.2—$FeO_3.H_2O$: Fe—62.9; O—27; H_2O—10.1
	Hematite: 5.1—Fe_2O_3: Fe—69.9; O—30.1
	Limonite: 3.8—$2Fe_2O_3.3H_2O$: Fe—59.8; O—25.7; H_2O—14.5
	Magnetite: 5.13—Fe_3O_4: Fe—72.4; O—27.6 (or FeO—31.0; Fe_2O_3—69.0)
	Pyrite: 5.03—FeS_2: Fe—46.6; S—53.4
	Pyrrhotite: 4.59—$Fe_{11}S_{12}$: Fe—61.5; S—38.5
	Siderite: 3.86—$FeCO_3$: FeO—62.1; CO_2—37.9
	Wolframite: 7.28—(Fe, Mn)WO_4: WO_3—76.5; (Fe, Mn)O—23.5
Copper	Azurite: 3.8—$Cu_3(CO_3)_2(OH)_2$: CuO—69.2; CO_2—25.6; H_2O—5.2
	Bornite: 5.15—Cu_5FeS_4: Cu—63.3; Fe—11.2; S—25.5
	Chalcanthite: 2.21—$CuSO_4.5H_2O$: CuO—31.8; SO_3—32.1; H_2O—36.1
	Chalcocite: 5.65—Cu_2S: Cu—79.8; S—20.2
	Chalcopyrite: 4.2—$CuFeS_2$: Cu—34.5; Fe—30.5; S—35.0
	Chrysocolla: 2.13—$CuSiO_3.2H_2O$: CuO—45.2; SiO_2—34.3; H_2O—20.5
	Covellite: 4.59—CuS: Cu—66.5; S—33.5
	Cuprite: 6.0—Cu_2O: Cu—88.8: O—11.2
	Enargite: 4.44—Cu_3AsS_4: Cu—48.3; As—19.1; S—32.6
	Malachite: 3.97—$(CuOH)_2CO_3$: CuO—71.9; CO_2—19.9; H_2O—8.2
	Tennantite: 4.64—$Cu_8As_2S_7$: Cu—57.5; As—17.0; S—25.5
	Tetrahedrite: 4.8—$Cu_8Sb_2S_7$: Cu—52.1; Sb—24.8; S—23.1
Nickel and Cobalt	Cobaltite: 6.07—CoAsS: Co—35.5; As—45.2; S—19.3
	Chloanthite: 6.5—$NiAs_2$: Ni—28.1; As—71.9
	Garnierite: 2.55—$H_2(Ni, Mg)SiO_4.xH_2O$: NiO—27.5; MgO—14.8; SiO_2—44.4; H_2O—13.3+
	Linnaeite: 4.9—Co_3S_4: Co—58; S—42
	Millerite: 5.53—NiS: Ni—64.7; S—35.3
	Niccolite: 7.49—NiAs: Ni—43.9; As—56.1
	Pentlandite: 4.75—(Fe, Ni)S_2: Ni—32.8; Fe—31.3; S—35.9
	Smaltite: 6.5—(Co, Ni)As_2: Co—22.0; Ni—21.9; As—56.1
Manganese	Manganite: 4.3—$Mn_2O_3.H_2O$: Mn—62.4; O—27.3; H_2O—10.3
	Psilomelane: 4.2—H_4MnO_5: H—2.9; Mn—39.5; O—57.6
	Pyrolusite: 4.78—$MnO_2.nH_2O$: Mn—63.2; O—36.8
	Rhodochrosite: 3.5—$MnCO_3$: MnO—61.7; CO_2—38.3
	Rhodonite: 3.56—$MnSiO_3$: MnO—54.1; SiO_2—45.9
Silver	Argentite: 7.29—Ag_2S: Ag—87.1; S—12.9
	Cerargyrite: 5.57—AgCl: Ag—75.3; Cl—24.7
	Dyscrasite: 9.64—Ag_3Sb: Ag—72.9; Sb—27.1
	Hessite: 8.41—Ag_2Te: Ag—62.9; Te—37.1
	Polybasite: 6.1—$(Ag, Cu)_9SbS_6$: Ag—75.6; Sb—9.4; S—15.0
	Proustite: 5.57—Ag_3AsS_3: Ag—65.4; As—15.2; S—19.4
	Pyrargyrite: 5.83—Ag_3SbS_3: Ag—59.9; Sb—22.3; S—17.8
	Stephanite: 6.25—Ag_5SbS_4: Ag—68.5; Sb—15.2; S—16.3
Lead	Anglesite: 6.25—$PbSO_4$: PbO—73.6; SO_3—26.4
	Cerussite: 6.51—$PbCO_3$: PbO—83.5; CO_2—16.5
	Galena: 7.5—PbS: Pb—86.6; S—13.4
	Wulfenite: 6.85—$PbMoO_4$: PbO—60.7; MoO_3—39.3
Zinc	Calamine: 3.45—$Zn_2SiO_4.H_2O$: ZnO—67.5; SiO_2—25.0; H_2O—7.5
	Smithsonite: 4.37—$ZnCO_3$: ZnO—64.8; CO_2—35.2
	Sphalerite: 4.02—ZnS: Zn—67.0; S—33.0
	Willemite: 4.04—Zn_2SiO_4: ZnO—73.0; SiO_2—27
	Zincite: 5.55—ZnO: Zn—80.3; O—19.7
Antimony Arsenic	Orpiment: 3.48—As_2S_3: As—61.0; S—39.0
	Realgar: 3.52—As_2S_2: As—70.1; S—29.9
	Stibnite: 4.57—Sb_2S_3: Sb—71.4; S—28.6
Titanium	Ilmenite: 4.7—$FeTiO_3$: Fe—36.8; Ti—31.6; O—31.6
	Rutile: 4.21—TiO_2: Ti—61.6; O—38.4 (Fe up to 10)
	Cassiterite: 6.95—SnO_2: Sn—78.6; O—21.4
	Cinnabar: 8.1—HgS: Hg—86.2; S—13.8
	Molybdenite: 4.75—MoS_2: Mo—60.0; S—40.0

As is true in many other instances, description is much slower and more tedious than manipulation, and the operations herein outlined may be performed easily in a few moments after one is accustomed to the use of the slide rule. To a person unfamiliar with the rule the question of placing the decimal point at once suggests itself. Used with the accompanying tables, the percentage values give all the assistance needed on this point and render the correct interpretation here more facile even than with the "geologist's slide rule." The values obtained should bear the same relations to each other as do the corresponding percentage values in the tables. For other operations the reader unaccustomed to slide-rule use is referred to various directions which are furnished by the manufacturers.

The accompanying tables are believed to be complete enough for most metamorphic, petrographic, and economic uses, but can be augmented to suit special needs. The percentages of a few oxides are included, as these are desirable in many cases. Any arrangement of minerals in such a table can be made to suit individual needs, and such tables mounted separately or back to back and covered with gelatin will be found an excellent form for field use, if such is desired.

LAKE SUPERIOR HIGHLANDS: THEIR ORIGIN AND AGE

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The dominant relief feature of the highland region about Lake Superior is an even, distinctly elevated plain in which the rivers are deeply intrenched. The genesis of this broad upland plain is a moot question of long standing. Its settlement involves the derivation and facial expression of the landscape over more than one-quarter of the entire North American continent.

At first glance the possibilities of the present peneplain's dating back in its formation to Early Cambrian, or even to pre-Cambrian, times, as is sometimes argued, seems so remote as almost to preclude serious consideration. Only by merest chance could there be survival of any remnant of so ancient a grade-plane. This chance lies in the exhuming of the old peneplain by the slow and uniform removal of a soft Paleozoic covering. Such a facet, if it persisted, would be quite small necessarily, coincidental, and, more properly, a product of some subsequent epoch of planation. Proofs of its antiquity have to be sustained by testimony overwhelmingly pertinent and convincing.

Indubitable evidence fixing the age of such a peneplain would not be likely to be found within the area of the highlands itself. It even might not be displayed anywhere within the limits of the great crystalline shield of Canada at all. Probabilities are for its disclosure far outside of the immediate elevated region. It is not a satisfactory solution of the problem to connect such an elevated and dissected plain with one of closely similar attitude emerging from beneath Cambrian sediments. In a continental geologic column there are many old plains—some only of provincial extent, but many of continental span or of dimensions as wide as such plains ever attain. To some place beyond the margins of the highland flats must attention first be turned for testimony bearing upon their age.

In casting about for a favorable locality in which to begin inquiry it is but natural to turn to some spot nearest to the highlands that displays a considerable section of the later geologic formations. There appears to be a section of this description off to the southwest of the middle angle of the Canadian shield, on the south side of the great tongue of pre-Cambrian rocks which extends from Lake Superior into northwestern Iowa where the old terranes are known as the Sioux quartzite.

On the general stratigraphic scheme of Iowa,[1] which is the latest and most complete of any now available, the erosion intervals may be readily indicated, together with the taxonomic values of each. Since some of these lines represent peneplanation sufficiently wide

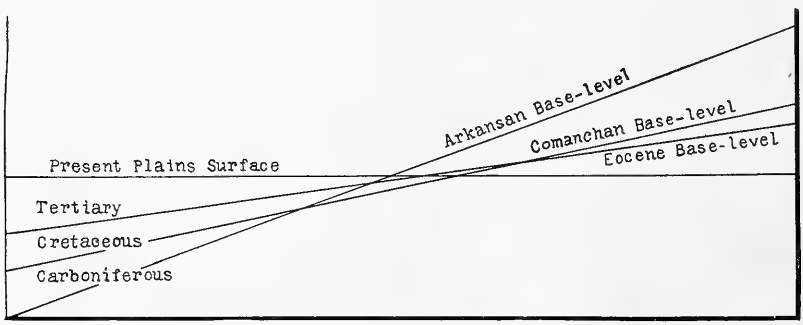

Fig. 1.—Relative attitudes of later base-levelings in Lake Superior region.

in extent to affect the Lake Superior region, they may be critically examined in turn, eliminated, or further compared in order that the parallelism with the highland plain under consideration may be exact.

Of the dozen or more conspicuous planes of unconformity which mark the geologic column of the region the Arkansan, Comanchean, and Eocene horizons are the most important. These represent most assuredly peneplains of great extent. Their intersections with the present ground-surface trend north-northwest through west-central Minnesota and Iowa. Along this line the intersections of all of them chance to be very close together. In diagram they may be represented as in Fig. 1.

[1] *Geol. Surv. Iowa*, XXII (1913), 154.

If the plane in which the lines of the diagram lie be considered as standing in a northeast-southwest direction the several peneplains of the Lake Superior region must have been successively obliterated, leaving only remnants of the very latest ones. The Arkansan peneplain, which constitutes the floor of the Coal Measures in the Upper Mississippi valley north of the Ozarks, may be for the present neglected. For reasons which are to follow, the Comanchean peneplain, which forms the Cretaceous floor in Iowa, Minnesota, and Manitoba, may be particularly examined.

Recent investigations about the point where the three states of Iowa, Minnesota, and South Dakota meet clearly disclose some instructive geologic structures bearing directly upon the problem under consideration. The field data thus acquired are supplemented by numerous deep-well records. By reference to the general geologic map of Iowa the Paleozoic formations are noticed to be distributed in relatively narrow belts trending in a northwest direction across the northeastern one-third of the state. Very singularly, it has always seemed, these belts abruptly terminate at the north soon after the state boundary is passed. This rather peculiar circumstance appears never to have excited curiosity as to its cause. Far to the north, about Winnipeg, in Manitoba, there is the same narrow belting of the same formations and, as farther south, the strike is northwest. The Canadian Paleozoic area is separated in central Minnesota from the Iowan Paleozoic field by a broad expanse of pre-Cambrian rocks.

Structurally these pre-Cambrian rocks form the core of a rather notable arch the axis of which runs northeast and southwest. This anticline is one of large proportions and extends from the east shore of Lake Superior to central South Dakota, where, as a canoe-shaped form, it plunges beneath the post-Paleozoic deposits of the Great Plains. The exposure of Sioux quartzite constitutes the western nose of the fold.

It is against the south slope of the sharp Siouan anticline that the belted Paleozoic terranes of northeastern Iowa are upturned, and there cut off. The eastern margin of the vast Cretaceous field crosses the same line, so that there is apparently no westward extension of the five groups of formations, if it ever existed, at

least on the surface of the ground. On the other, or north, side of the anticline the same belts recur, as already stated.

Bearing in mind the geographic position of this marked anticline, an arch between the center of which and the bases of its limbs there is a stratigraphic interval of more than 5,000 feet, it is quite obvious that the Paleozoic belts originally did not really terminate against it in southern Minnesota but rather extended over it and were continuous with the similar Canadian belts. This being the case, it is equally obvious that the Iowan belts should not only not stop against the arch, but should continue westward along the strike of the fold, but beneath the Cretaceous covering. This is found actually to accord with recently observed facts. A cross-section (Fig. 2), which is drawn to scale, indicates the actual amount of tilting displayed at the present time, with the part originally laid down, but removed during Mid-Cretaceous times, represented by dotted lines.

There are incontrovertible proofs fixing within very narrow limits the geologic date of the uprising of the great Siouan arch, and also the time of its complete reduction again to an even plain lying but little above the level of the sea. Since all of the Paleozoic formations, from Cambrian to latest Carboniferous, take part in the folding while the Cretaceous strata do not, it is manifest that the main movement occurred in Early Mesozoic times, largely during the Triassic and Jurassic periods. Comanchean time (Early Cretaceous) in the region must have been principally a period of rapid, enormous, and very complete denudation, since by the beginning of Mid Cretaceous time, when marine deposition over this part of the continent took place, the floor upon which the sediments were laid down was as even as any known peneplain. The Cretaceous floor is a true plain worn out on the beveled edges of the Paleozoics and older rocks.

The areal distribution and attitude of the Cretaceous beds along their eastern margin is suggestive. In Minnesota the western half of the state is occupied by deposits of Cretaceous age. Outliers occur far to the eastward—to the Mississippi River and the Mesabi Range. The great thickness of Cretaceous deposits in the eastward-facing escarpment of the Duck and Riding mountains, in

Manitoba, and their gentle dip to the west point to their former great eastward extension probably to Lake Superior and Hudson Bay. There is little reason to doubt, therefore, the assumption that the Cretaceous sediments once covered the Lake Superior

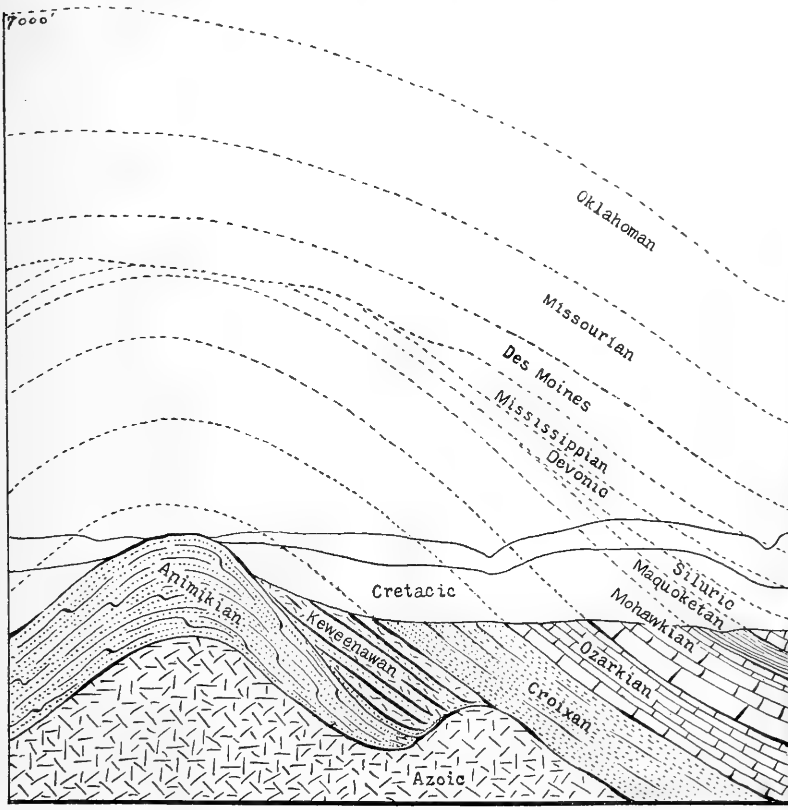

FIG. 2.—Erosion of the Triassic Siouan mountains in southern Minnesota.

highlands and that the peneplain which formed the Cretaceous floor must have imparted the main topographic expression to the highland plain.

In Iowa, and Minnesota also, the Cretaceous strata themselves are notably beveled, but at a lower angle than that of their floor. There are good reasons for believing that this Tertiary base-leveling took place in Eocene time. There appear to be important water-

laid deposits of this age in the region; but Miocene and Pliocene deposits seem to be chiefly, if not entirely, continental in character.

It is quite probable, therefore, that the tops of the monadnocks which rise above the level of the highland plain of Lake Superior represent approximately the level of the Comanchean peneplain; and that the present plain of that region is Eocene in date. This interpretation has many points supporting it; and but few or none against it. Around Lake Superior glacial action has entirely removed the evidence for the ready determination of these supporting facts. The strongest and most convincing testimony comes from neighboring localities; and the several lines of evidence appear to be wholly congruous and mutually supporting.

For the formation of the Lake Superior highlands an Eocene age appears to be conclusive. The necessary consequences are far-reaching. It calls for an examination anew of the geographic development of a large part of the northeastern section of the North American continent.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA FOR 1909, 1910, 1911, AND PART OF 1912

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V. QUEBEC

Barlow[1] states that the Chibougaman region of Quebec is underlaid by Keewatin, Laurentian, and Huronian formations, and a gabbro anorthosite.

Barlow[2] states that the Chibougaman region, lying between latitudes 49° 30′ and 50° 11′ 06″ north and longitude 70° 03′ 40″ to 74° 35′ 50″, about 300 miles north of Montreal, has the following pre-Cambrian succession region, beginning with the oldest: (1) Keewatin, serpentine and pyroxenite; deformed quartz porphyry and porphyrites, basalts, diabases and gabbros; (2) anorthosite intrusive into Keewatin; (3) Laurentian granites, granite and diorite gneiss intrusive into (1) and (2); (4) unconformity; (5) Lower Huronian—slates, sandstones, arkose and conglomerate.

Bancroft[3] classifies the pre-Cambrian rocks of the Harricanaw and Nottaway rivers of northwestern Quebec as Keewatin metamorphosed extrusive and intrusive basic and acid igneous rocks with bands of highly altered sedimentary rocks; Laurentian granites and granite gneisses intrusive into the Keewatin; Lower Huronian (?) conglomerates, graywacke, slate, and arkose, injected by granite; post-Middle Huronian (Keweenawan?) dikes of quartz and olivine diabase.

[1] "Geology of the Chibougaman Region, Quebec," *Canada Geol. Soc. Am. Bull.*, XXII, No. 4 (1911), 738 pp.

[2] "Geology and Mineral Resources of the Chibougaman Region," Report by the Chibougaman Mining Concession, *Province of Quebec, Canada*, 1911, 215 pp., 68 pls., 19 figs., 2 maps.

[3] J. A. Bancroft, "Geology and Natural Resources of the Basins of Harricanaw and Nottaway Rivers, Northwestern Quebec," *Province of Quebec*, 1912, 16 pp., 1 map.

Cirkel[1] states the Amherst graphite deposits occur as lenticular masses and pockets, disseminations, and veins in eruptive rock intrusive into the Grenville limestones. Graphite occurs within feldspar, quartz, and pyroxene. It is mostly associated with feldspar or pyroxene, less frequently with calcite and wollastonite. Garnet, zircon, scapolite, titanite, muscovite, pyrite, apatite, leucoxene, biotite and monazite are accessory minerals of the gangue.

Dulieux[2] reports that the rocks of the Lake Chibŏgomon region, north of Lake St. John, in Quebec, constitute an isolated belt of diabases and gabbros and rocks derived from them; viz., chloritic, epidotic, talc schists, and serpentine, surrounded by Laurentian granites and gneisses, the backbone of Labrador. They contain deposits of cupriferous and auriferous sulphides, magnetic iron, and asbestos.

Wilson[3] finds that the pre-Cambrian rocks in western Quebec between 75° 30′ and 79° 30′ west longitude and 48° and 49° north latitude include hornblende schists, chlorite schists, diabase, and porphyries, all probably of Keewatin age, and Laurentian granites and gneisses.

VI. THE CORDILLERAS

Ball[4] finds that the pre-Cambrian, magnetic, titaniferous iron ores of Iron Mountain in southeastern Wyoming form a series of dikes cutting anorthosite. Granitic masses dissect both. The ores and the anorthosite contain the same minerals with some minor exception, but in different proportions. The tonnage of ore is large, averaging high in iron, and probably low in phosphorus, but too high in titanium to be desirable under present conditions.

[1] Fritz Cirkel, "The Amherst (Quebec) Graphite Deposits," *Quart. Bull. Can. Mining Inst.*, 1910, pp. 107–15.

[2] E. Dulieux, "The Chibŏgomon Region, Quebec," *Jour. Can. Min. Inst.*, XII (1909), 184–93.

[3] W. J. Wilson, "Geological Reconnaissance along the Line of the National Transcontinental Railway in Western Canada," *Canada Geol. Survey, Mem.* 4, (1910), pp. 56, 5 pls., 1 geol. map.

[4] Sidney H. Ball, "Titaniferous Iron Ore of Iron Mountain, Wyoming," *Bull. 315*, 1906, pp. 206–12.

Bancroft[1] states that the oldest rocks in Northern Yuma County, Arizona, are granites and diorites, probably of Archaean age. They are capped by a thick series of metamorphosed pre-Cambrian sediments, quartzites, limestones, dolomites, argillites, arenaceous shales, and quartz mica schists, interlayered and intruded by basic rocks altered to amphibolites.

Blackwelder[2] reports that the pre-Cambrian of three geologic sections in western Wyoming consists of gneiss, gray, and red granites penetrated by diabase and other dikes, hornblende gneiss, and granite with large dikes of diabase.

Blackwelder[3] states that the pre-Cambrian rocks of the Laramie and Sherman quadrangles of northeastern Wyoming include: (1) hornblende schists, associated with some schistose rhyolites and felsites, soft mica gneisses et cetera, and rocks resembling contorted quartzites, and limestones, all highly metamorphosed; these rocks are the oldest in the area; (2) mildly metamorphosed granite gneiss representing two periods of intrusion, both younger than (1); (3) basic intrusives, comprising syenites, gabbros, diorites, granodiorites, and gabbro gneiss, all younger than (2); (4) granite porphyry intrusive into (1) and (2); (5) anorthosite, intrusive into (1), (2), and the gabbro gneiss of (3); (6) granite intrusive into (1), (2), (3), (4), and (5); (7) small diabasic and dioritic dikes injected into (6) and probably younger than (7).

Blackwelder[4] finds an unconformity in Big Cottonwood Canyon along the upper course of the Ogden River in the Wasatch Mountains, between slightly fossiliferous early Cambrian quartzites and an older quartzite slate series, several thousand feet thick, which he assigns to the Algonkian in accordance with the prevailing methods of correlation.

[1] Howland Bancroft, "Reconnaissance of the Ore Deposits in Northern Yuma County, Arizona," *Bull.* 451, *U.S. Geol. Survey*, 1911, pp. 130, maps and illustrations.

[2] Eliot Blackwelder, "A Reconnaissance of the Phosphate Deposits in Western Wyoming," *Bull. 470, U.S. Geol. Survey*, 1911, pp. 452–81.

[3] N. H. Darton, E. Blackwelder, C. E. Siebenthal, "The Laramie and Sherman Quadrangles, Wyoming," *U.S. Geol. Survey, Folio No. 173*, 17 pp., 8 pls., 3 figs.

[4] Eliot Blackwelder, "New Light on the Geology of the Wasatch Mountains, Utah," *Bull. Geol. Soc. Am.*, XXI (1910), 517–42, pls. 36–40.

The unconformity is shown by the variable thickness of the quartzite-slate series, slight discordance of bedding, and a basal conglomerate.

The Algonkian system consists of alternating quartzites, slates, and conglomerates, which are variable from place to place; and show cross-bedding, ripple marks, and mud cracks. The materials are not well assorted, and in the sandy beds, the prevailing colors are shades of red; green predominates in the shales. Fossils and limestone appear to be lacking. Blackwelder is inclined to view these sediments as subaerial, largely fluviatile deposits.

Cross and Hole[1] state that the pre-Cambrian in the Engineer Mountain quadrangle of Colorado include Archaean gneisses and schists derived from granites, diorites, and diabase. These are intruded by granitic and gabbroic masses. Several thousand feet of massive and some thin-bedded quartzites, white, brown, and black, and occasional bands of brown to black shale, all of Algonkian age, overlie the Archaean.

Diller[2] describes the occurrence of asbestos in the pre-Cambrian rocks near Casper, Wyo., and the Grand Canyon of Arizona. At Casper, Wyo., the asbestos occurs mostly at the igneous contact between serpentine and granite. In the Grand Canyon, asbestos occurs only where diabase sills have been intruded into Algonkian magnesian limestones.

Diller[3] states that the rocks of the asbestos area south and southeast of Casper, Wyo., are hornblende schist, diorite, granite, and serpentine of pre-Cambrian age.

In the Grand Canyon of Arizona, certain Algonkian limestone beds contain asbestos where they have been intruded by diabase sills.

Daly[4] finds that the great stratigraphic problem of the Shuswap Lakes region is the correlation of the non-fossiliferous, crystalline

[1] W. Cross and A. H. Hole, "Engineer Mountain Folio," *U.S.G.S. Folio* 171, 1910, 14 pp., 5 pls., 1 fig.

[2] J. S. Diller, "The Types and Modes of Occurrence of Asbestos in the United States," *Jour. Can. Min. Inst.*, 1911, pp. 93–107.

[3] J. S. Diller, "The Types, Modes of Occurrence, and Important Deposits of Asbestos in the United States," *Bull. 470, U.S. Geological Survey*, 1911, pp. 505–23.

[4] R. A. Daly, "Reconnaissance of the Shuswap Lakes and Vicinity (South Central British Columbia), *Canada Department of Mines*, Ottawa, 1912.

rocks which lie unconformably beneath the Carboniferous. Dawson found that an unconformity separated them into two divisions. The rocks beneath this unconformity he classed as pre-Cambrian and those above as Cambrian. Daly classifies all the rocks unconformably beneath the Carboniferous as pre-Cambrian, excepting a part of the quartzite rocks immediately underlying the Carboniferous which he believes may be Cambrian. The unconformity which Dawson believed separated the Cambrian from the pre-Cambrian Daly places within the pre-Cambrian, and calls the rocks below this unconformity pre-Beltian, and those above it Beltian, with the reservation previously stated that a part of the upper portion may be Cambrian.

The pre-Beltian consists of the dominantly Sedimentary Shuswap series, about 26,500 feet thick, which is intruded by batholiths, dikes, and sills of granite. It includes limestones, metargillites, mica-schists, paragneisses, green schists and greenstones, the latter probably of extrusive origin although no definite extrusive characteristics have been found. The Beltian is also dominantly sedimentary, consisting from the base up of about 200 feet of arkose sandstone overlain by about 18,000 feet of argillaceous rocks interbedded with limestones and quartzites, the whole constituting the Albert Canyon division. Above this is the Glacier quartzite division about 25,000 feet thick, the upper part of which, Daly believes, may be Cambrian, although no fossils have been found.

The strike of the pre-Cambrian rocks is nearly east and west or practically at right angles to the Cordilleran folding. Normal faulting in part contemporaneous with deposition is more evident than folding, and the bedding is either nearly horizontal or only slightly inclined. From the fact that both the Shuswap and the Beltian are dominantly schistose with their schistosity essentially parallel with the slightly deformed bedding. Daly infers that the schistosity has not resulted from tangential stress but from the weight of the sediments and the action of thermal solutions. While the pre-Cambrian rocks are only gently inclined, the Carboniferous and Triassic rocks of this region are highly folded.

The rocks of the Rocky Mountain geosynclinal are believed by Daly to be largely derived from the pre-Beltian. He finds that the microscopic characteristics of the feldspars in the sediments are identical with those of the granites, sills, and dikes which intrude the Shuswap series. The extraordinary fine grain of the quartzites he attributes to their derivations from metargillites, phyllites, and greenstone schists.

Daly does not care to have the term Algonkian applied to the pre-Cambrian of the Shuswap Lakes region, since it would include two series which are separated by a well marked unconformity.

Darton[1] records a few observations on the distribution of the Vishnu and Grand Canyon series in western New Mexico and northern Arizona.

Emmons[2] states that a complex of pre-Cambrian granites constitutes the basement upon which later sedimentary and igneous rocks of the Cananea district were superimposed.

Emmons[3] states that the oldest rocks of the Little Rocky Mountains between the Missouri and Milk rivers consist of pre-Cambrian quartzite by intrusive porphyry.

George and Crawford[4] find that the pre-Cambrian rocks of the Hahn's Peak region of Routt County, Colo., as a complex of schists, gneiss, gneissoid granite, and granite with isolated patches of quartzite.

Henderson[5] describes the pre-Cambrian rocks of the Front Range of Colorado, as consisting of Archaean granite and gneiss, and an Algonkian quartzite, outcropping from South Boulder Canyon

[1] N. H. Darton, "A Reconnaissance of Parts of Western New Mexico and Northern Arizona," *Bull 435, U.S. Geological Survey*, 1910, pp. 14–15.

[2] S. F. Emmons, "Cananea Mining District of Sonora," *Econ. Geol.*, V, No. 4 (1910), 312–.

[3] W. H. Emmons, "Gold Deposits of the Little Rocky Mountains, Montana," *Bull. 340, U.S. Geological Survey*, 1907, pp. 96–116.

[4] R. D. George and R. D. Crawford, "The Hahn's Peak Region, Routt County, Colorado," *Colo., Geol. Surv., 1st Rept.*, 1908, pp. 189–229, 1 pl., 1909.

[5] Junius Henderson, "The Foothills Formations of Northern Colorado," *Colo. Geol. Surv., 1st Rept.*, pp. 149–85, 1909.

southward a distance of about twelve miles. The latter is mostly white and in places coarsely conglomeratic, showing both bedding and cross-bedding.

Lindgren, Groton, and Gordon[1] find small areas of pre-Cambrian rocks, most of them in the northern part of New Mexico, exposed in eroded regions of doming and faulting, flanked by younger rocks. The largest, about 120 miles long, extends as a southward continuation of the Sangre de Cristo range of Colorado, to about 20 miles south of Santa Fe. It consists of a sedimentary series, chiefly quartzitic, which with certain greenstone tuffs, amphibolites, and rhyolites has been invaded by powerful intrusions of normal, usually reddish microcline granite, locally schistose. The granite in turn has been intruded by aplite, abundant pegmatite, and some masses or dikes of greenstone. The sediments are believed to correspond to those of Colorado similarly imbedded in red granite and to the quartzitic schists of the Pinal formation of Arizona, but the pre-Cambrian sediments of the Grand Canyon are believed to be younger than the red granite intrusions.

Unimportant ore deposits of gold and silver, some zinc and copper, but little lead are found in the granites, gneisses, and amphibolitic schists. Some rich gold placers have developed from them. The ores occur in quartz-filled fissures, usually of the "lenticular" type, shear zones filled with quartz stringers, and disseminations of sulphides in amphibolitic schists. The deposits frequently show evidence of deep-seated origin, such as the development of heavy minerals, and massive, glassy vein quartz. Many of them were folded and fractured. The ore deposition followed granitic and other intrusions, and the metals were probably delivered from the magmas.

Noble[2] describes the pre-Cambrian succession of the Shimuno area of the Grand Canyon. The following is a summarized section:

[1] W. Lindgren, L. C. Groton, and C. H. Gordon, "The Ore Deposits of New Mexico," *Prof. Paper 68, U.S. Geological Survey*, 1910, pp. —, several maps.

[2] L. F. Noble, "Contributions to the Geology of the Grand Canyon, Arizona. The Geology of the Shimuno Area, *Am. Jour. Science*, 4th Series, XXIX, 369–86.

UNCONFORMITY AT BASE OF CAMBRIAN

	Micaceous shaly sandstone uniform in composition but varying in color. Ripple marked and cross-bedded throughout. 2,297 feet thick.
	Basal conglomerate, 6 feet
Unkar Group. Thickness, 478*a* feet.	Calcareous shales and dolomitic limestones and dolomites containing some ripple marks, sun-cracks; total thickness, 304 feet.
	Ripple-marked, cross-bedded jaspers, quartzitic jasper, intrusive diabase, blue, red, vermilion, argillaceous and arenaceous shales. Thickness 579 feet.
	Pure fine-grained sandstones and quartzites, ripple-marked and cross-bedded. 1,564 feet thick.
Unconformity.	
Vishnu series.	Quartz schist, quartz mica schist, quartz hornblende schist, hornblende schist, quartz diorite and pegmatite granite.

Paige[1] states that the pre-Cambrian complex of Grant County, New Mexico, comprises various types of coarse-to-medium-grained granites and porphyries, and pegmatitic dikes.

Patton[2] finds that the pre-Cambrian rocks of the Grayback Mining district of Costilla Co., Colorado, consist of granite, granite and biotite gneiss, and a small amount of hornblende schist, amphibolite, and pegmatite.

Pardee[3] describes the sedimentary rocks of the Upper St. Joe River in Idaho which are supposed to be equivalent to the Algonkian formations of the Coeur d'Alene district. They consist of sandstones, graywackes, quartzites, shales, garnetiferous mica schists, and limestone.

Ransome[4] states that the basal rocks of the Breckenridge district

[1] Sidney Paige, "Metalliferous Ore Deposits near the Burro Mountains, Grant County, New Mexico," *Bull. 470*, 1911, *U.S. Geological Survey* (1911), pp. 131–50, 1 map.

[2] H. B. Patton et al., "Geology of the Grayback Mining District," *Colo. Geological Survey, Bull., 2*, 1910, 111 pp., 9 pls.

[3] J. T. Pardee, "Geology and Mineralization of the Upper St. Joe River, Idaho," *Bull. 470*, 1910, pp. 39–61, 1 map.

[4] F. L. Ransome, "Geology and Ore Deposits of the Breckenridge District," *U.S.G.S., Prof. Paper 75*, 1911, pp. 187.

of Colorado appear as isolated patches of pre-Cambrian schists and gneisses cut by pegmatite and granite.

Schrader[1] states that the principal topographic features of Mohave County, Arizona, are barren, plainlike, detritus-filled valleys 10 or 12 miles in width, separated by ranges of about equal width. Their trend is a little west of north, and the general slope is toward the southwest.

The ridges consist mainly of pre-Cambrian granites, porphyries, gneisses and schists cut by pre-Tertiary intrusives. The pre-Cambrian probably was once covered or nearly covered by Paleozoic sediments which are now almost completely eroded.

Schrader[2] finds that the pre-Cambrian granites, gneisses, and schists constitute the basement complex on which later sediments of Mohave County, Arizona, have been deposited. No pre-Cambrian sediments have been recognized. The trend of the schists is about N. 30° E. and their dip is either vertical or steeply east-northeast.

Smith[3] believes that the older gneisses of the Sierra Nevada are probably Archaean.

Spurr[4] states that the lowest formation in the Aspen district is pre-Cambrian granite.

Stone[5] finds that in the southeastern part of T. 5 N., R. 2 W., of Montana there is a flat area, underlain by Algonkian sediments of unknown thickness, mostly red shales, including some green shale and thin sandstone beds. They are believed to correspond to the Spokane shale to the east.

[1] F. C. Schrader, "The Mineral Deposits of Mohave County, Arizona," *Bull. 340, U.S. Geological Survey*, 1907, pp. 53–95.

[2] F. C. Schrader, "Mineral Deposits of the Cerbat Range, Black Mountains, and Grand Wash Cliffs, Mohave County, Arizona," *Bull. 397, U.S. Geological Survey*, 1909, pp. 220.

[3] James Perrin Smith, "The Geological Record of California," *Jour. Geol.*, XVIII, No. 3 (1910), 216–27.

[4] J. Edward Spurr, "Ore Deposition at Aspen, Colorado," *Econ. Geol.*, IV (1909), 301–20.

[5] R. W. Stone, "Geologic Relations of Ore Deposits in the Elkhorn Mountains, Montana," *Bull. 470, U.S. Geological Survey*, 1911, pp. 75–98, 1 map.

Turner[1] describes the complex of pre-Cambrian granite-gneiss, quartz-monzonite gneiss, granite augen schists, calcareous augen schists, and small lenses of hydrous mica-schists of the Silver Peak district of Nevada. Certain dolomites, quartzites, and green knotted schists underlying the Olenellus fauna may be Algonkian, but have been provisionally referred to the Lower Cambrian.

Walcott[2] finds the following succession of pre-Cambrian sediments exposed below basal Cambrian conglomerate in the Bow River valley of Alberta, Canada: Hector formation—shales 1,300–2,150 feet; Corral Creek formation—sandstones 2,600 feet, base not found. He correlates them with the Camp Creek and Sheppard series of arenaceous rocks which lie beneath the Cambrian and above the Siyet limestone in Montana, southwestern Alberta, and southeastern British Columbia.

Weeks[3] finds that the oldest rocks of the Osceola district are granites and schists, with intruded porphyry, which he believes are probably Archaean.

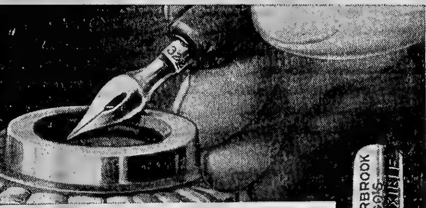
Winchell[4] finds that the lowest rocks exposed in the Ruby range, about 60 miles south of Butte, Mont., are quartz schists and slates, probably of pre-Cambrian age.

[1] H. W. Turner, "Geology of the Silver Peak Quadrangle," *Bull. of Geol. Soc. Am.*, XX (1909), pp. 223–64, 4 pls., 1 fig.

[2] C. D. Walcott, "Pre-Cambrian Rocks of the Bow River Valley, Alberta, Canada," *Smith. Misc. Coll.*, LIII, No. 7 (August, 1910), pp. 423–31, 3 pls.

[3] F. B. Weeks, "Geology and Mineral Resources of the Osceola Mining District, White Pine County, Nevada," *Bull. 340, U.S. Geological Survey*, 1907, pp. 117–33.

[4] Alexander N. Winchell, "Graphite near Dillon, Montana," *Bull. 470, U.S. Geological Survey*, 1911, pp. 528–32.



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REVISION OF PRE-CAMBRIAN CLASSIFICATION IN ONTARIO[1]

WILLET G. MILLER AND CYRIL W. KNIGHT
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The pre-Cambrian rocks of Ontario have undergone much scrutiny during the last seventy years. At the very beginning of his labors as provincial geologist of Canada (Ontario and Quebec) in 1843 Logan confirmed the observations made by other investigators, that the oldest fossiliferous rocks lie unconformably on a vast crystalline series. By degrees he and his associates, working in the regions adjacent to the Ottawa River and to the north shore of Lake Huron respectively, came to the conclusion that these ancient rocks as regards their age relations are divisible into two great groups, the older of which was named the Laurentian and the younger the Huronian.[2] Other pre-Cambrian rocks were studied on Lake Superior.[3] To these was given the name Upper Copper-bearing Rocks of Lake Superior. They were divided into two groups, the upper of which has come to be known as the Keweenawan and the lower as the Animikie.

[1] This paper was presented in abstract at the meeting of the Geological Society of America, December, 1914.

[2] *American Association for the Advancement of Science*, 1857; *Geology of Canada*, 1863, pp. 22 f.

[3] *Geology of Canada*, 1863, pp. 67 f.

In conformity with the views of the time, Logan and his associates considered the Laurentian to be made up essentially of metamorphosed sedimentary material. With the passing of the years, the Laurentian came to be looked upon as separable into

FIG. 1.—Geological sketch map of Ontario

three groups: (1) Lower Laurentian or Ottawa series, consisting essentially of granite gneiss; (2) Middle Laurentian or Grenville series, consisting of crystalline limestone and other rocks; and (3) Upper Laurentian or Norian. From the Laurentian groups was first taken the Upper Laurentian, which was proved to be not of sedimentary origin but an intrusive.[1] Afterward the Grenville

[1] *Neues Jahrb.*, Beilage Band 8, 1893, pp. 419–98.

was taken away and the name Laurentian came to stand only for granite and gneiss of undoubted igneous origin, older than the so-called Huronian sediments.[1]

CLASSIFICATIONS NOW IN USE

The need for a revision of the classification and nomenclature of the pre-Cambrian rocks has been felt from time to time as knowledge concerning them has increased.

Table I shows the classification that has been adopted by the authors.[2] Like all other classifications it will be subject to revision when a more perfect understanding of the relations of the rocks is acquired.

In Table II are given the classifications employed by several other authors.

LAURENTIAN AND HURONIAN

Logan, himself, when introducing the names Laurentian and Huronian said: "These local names are, of course, only provisional, devised for the purpose of avoiding paraphrastic or descriptive titles, the use of which had been found inconvenient, and they can be changed when more important developments, proved to be the equivalents of the series, are met with elsewhere."[3] Hence, when it is found advisable to change the classification or to discard the now historic names, looked on by Logan as being only provisional, mere sentiment should not be permitted to stand in the way. In most cases it is better to discard a name rather than employ it in a new sense.

As shown in a preceding paragraph, the term Laurentian is now used in a much-restricted sense and, as will be seen from the following pages, different meanings are attached to it by various authors. The earlier literature in which Laurentian rocks were described has become almost unintelligible.

Much confusion has arisen also through the employment of the classic name Huronian, especially with the prefixes Upper, Middle, and Lower, in different senses. The term Lower Huronian, for example, has been applied indiscriminately to certain rocks

[1] Report of International Committee, *Jour. Geol.*, February-March, 1905.

[2] *Ont. Bur. Mines*, XXII, Part II, Appendix.

[3] *American Association for the Advancement of Science*, 1857.

that lie below one of the greatest known unconformities, that between the Timiskamian and Animikean of our classification, as well as to those above it. Again, rocks have been classed as Upper Huronian, or Animikie, which have not been proved to be

TABLE I

PRE-CAMBRIAN

KEWEENAWAN	
Unconformity	
ANIMIKEAN	Under this heading the authors place not only the rocks that have heretofore been called Animikie, but the so-called Huronian rocks of the "classic" Lake Huron area, and the Cobalt and Ramsay Lake series. Minor unconformities occur within the Animikean.
Great unconformity	
(ALGOMAN granite and gneiss) *Igneous contact*	Laurentian of some authors, and the Lorrain granite of Cobalt, and the Killarney granite of Lake Huron, etc.
TIMISKAMIAN	In this group the authors place sedimentary rocks of various localities, that heretofore have been called Huronian, and the Sudbury series of Coleman.
Great unconformity	There is no evidence that this unconformity is of lesser magnitude than that beneath the Animikean.
(LAURENTIAN granite and gneiss) *Igneous contact*	
LOGANIAN { Grenville (*Sedimentary*) Keewatin (*Igneous*)	The authors have found the Keewatin to occur in considerable volume in southeastern Ontario and have determined the relations of the Grenville to it.

different in age from certain so-called Lower and Middle Huronian. It may be added that for many years the Keewatin also was called Lower Huronian. Even at the present time when making use of the name Huronian, in order to secure clearness, it is necessary to say in what sense the name is employed, whether in that of the United States Geological Survey or in that of Coleman or Lawson

et al. Owing to the confusion that has arisen in connection with the use of the term the authors think it well to discard it.

Among those, including Coleman, Collins, and the authors, who have studied during recent years the pre-Cambrian rocks of the Lake Huron–Lake Timiskaming region, there is agreement as to most of the age relations, as will be seen from the classifications on a following page, but different nomenclatures are employed.

As to facts, the chief difference in opinion between Coleman and the authors is in regard to the relations among what he calls the Lower and Upper Huronian and the Animikie, and what we call collectively the Animikean. It is agreed that the Cobalt series, the Ramsay Lake series of Sudbury, and the usually flat-lying sediments of the classic area of Lake Huron are of the same age. But in Coleman's opinion, as shown in his table, the sediments of the Sudbury basin, the so-called Animikie series, are younger than the Ramsay Lake series, which he classes as Lower Huronian. But no discordance has been proved to exist between the Ramsay Lake series and the Animikie of the near-by Sudbury basin. For this and other reasons the authors class the Ramsay Lake series and the Animikie together as the Animikean. In this they are supported by investigations made by Van Hise and Seaman, who say: "About half a mile east of Sudbury they found a breccia or conglomerate [Ramsay Lake series] similar in every way to that seen at Onaping [Animikie series of Sudbury basin] resting on the eroded edges of Huronian [Timiskamian] and containing fragments of it."[1]

Beginning with the bottom of the columns of the two classifications, it will be seen that Coleman has accepted, provisionally at least, our interpretation of the relations of the Keewatin and the Grenville, described in a recent report.[2]

As is shown by Table II, Coleman and Collins use the name Sudbury series or Sudburian for the rocks which we call Timiskamian. But Timiskamian has priority as it was employed by the Ontario Bureau of Mines for rocks that occur at various points over an area at least 5,000 square miles in extent before the term "Sudburian" was introduced.[3] Moreover, since Logan first

[1] *U.S. Geol. Survey Bull.* 360, p. 425. [2] *Ont. Bur. Mines*, XXII, Part II.

[3] *Eng. and Min. Jour.*, September 30, 1911, p. 648.

studied his "Huronian" rocks on Lake Timiskaming, the name Timiskamian is appropriate for a part of them. Further, the name Sudburian is likely to be confused with sudburite, a name applied to certain basic igneous rocks at Sudbury that intrude the Timiskamian but are older than the Algoman granite and gneiss.[1] It may be added that basic rocks of the age of sudburite are widespread and have frequently been mistaken for Keewatin. They are represented by the peridotite and augite lamprophyre of the Porcupine region, one hundred miles north of Sudbury, and by the lamprophyres and other rocks of Cobalt, ninety miles to the northeast of Sudbury. For economic purposes, at least, some age name should probably be applied to these basic rocks. Sudburian would be suitable had it not been used in another sense.

While Coleman and the authors are agreed on the age relations of the two great series of granites and gneisses, different nomenclatures are employed. The rocks that he calls "Granites Eruptive through the Keewatin," and to which he gives no distinctive age name, we class as Laurentian. In his classification Laurentian is applied to the granites and gneisses that we, following Lawson's nomenclature, class as Algoman.

The appellation Laurentian was first given to the rocks of the district bordering the Lower Ottawa River, and the older granites and gneisses here are different in age from the Algoman. The fact that the rocks now called Algoman have frequently been classed as Laurentian should not be given consideration. If the name Laurentian is to be retained, it should be applied to the oldest granites and gneisses. The following quotation shows that Logan did not intend that the name should be applied to granites that are intrusive into any part of his "Huronian." Further, it will be seen that, in the classic area of Lake Huron, Logan recognized both the Laurentian which, in his opinion, underlies all of his Huronian sediments, including that part of them now known as Timiskamian, and granites eruptive into the Huronian. Unfortunately later work has shown that much, at least, of what he considered to be Laurentian is really the intrusive granite which he called Huronian granite.

[1] *Ont. Bur. Mines*, XXIII, 215–17.

The intrusive granite occupies a considerable area on the coast of Lake Huron, south of Lake Pakokagaming. It there breaks through and disturbs the gneiss of the Laurentian series and forms a nucleus from which emanates a complexity of dykes, proceeding to considerable distances. As dykes of a similar character are met with intersecting the rocks of the Huronian series [Timiskamian], the nucleus in question is supposed to be of the Huronian age, as well as the greenstone dykes which are intersected by it.[1]

If Logan mistakenly applied the name Laurentian to his "Huronian granite" in certain areas, why should his mistake be perpetuated? Now that we know these granites and gneisses in some localities belong to what he called Huronian granite, why should we apply the name Laurentian to them instead of to the group that is much older, the Laurentian of our classification and the "Granites Eruptive through the Keewatin" of Coleman's?

There are certain objections to giving the names of systems or series to intrusives such as the Laurentian and Algoman. For this reason the authors prefer to inclose the names in parentheses in the table of classification, thus: (Laurentian intrusives) and (Algoman intrusives), or (Laurentian) and (Algoman). These intrusives occur in such great volume in the pre-Cambrian that it would seem well to retain distinctive age names for them. Age names are especially useful in economic work, our most important gold deposits, for example, being genetically connected with the Algoman.

DUAL SUBDIVISION

A dual subdivision of the pre-Cambrian, as shown in Table II, is employed by many authors. It is purely arbitrary, and is now known to be based on a misconception as to the relations of the rocks. It was believed, or assumed, that the greatest unconformity within the pre-Cambrian is that at the base of the "Lower Huronian," as the term is used by the U.S. Geological Survey. Such, however, has not proved to be the case. There is no proof that the unconformity at the base of the Timiskamian is of less magnitude than that at the base of the Animikean, of the authors' classification, or vice versa. According to Coleman, the various sedimentary series of the Timiskamian, his Sudbury series, where

[1] *Geology of Canada*, 1863, p. 58.

TABLE II

AGE CLASSIFICATIONS OF PRE-CAMBRIAN ROCKS OF ONTARIO PROPOSED BY VARIOUS AUTHORS

Miller and Knight*		Coleman†		Collins‡		Lawson§		*U.S. Geol. Survey Bull. 360*‖
KEWEENAWAN	POST LAURENTIAN	KEWEENAWAN	HURONIAN	KEWEENAWAN	ALGONKIAN	KEWEENAWAN (Nipigon)	ALGONKIAN	KEWEENAWAN
Unconformity		*Unconformity*		*Unconformity*		*Unconformity*		*Unconformity*
ANIMIKEAN (with minor unconformities)		ANIMIKIE *Unconformity* UPPER HURONIAN *Small unconformity* LOWER HURONIAN		WHITEWATER LORRAIN *Local unconformity* COBALT		ANIMIKIE		UPPER HURONIAN *Unconformity* MIDDLE HURONIAN *Unconformity* LOWER HURONIAN
Great unconformity		*Great unconformity*		*Great unconformity*		*Eparchean interval*		*Unconformity*
(ALGOMAN INTRUSIVES)		(LAURENTIAN, in eruptive contact with lower rocks)	PRE-HURONIAN	Batholithic granite intrusives	ARCHEAN	ALGOMAN (granite-gneiss, batholithic in HURONIAN)		
Intrusive contact				*Intrusive contact*		*Irruptive contact*		
TIMISKAMIAN	PRE-LAURENTIAN	SUDBURY		SUDBURY, TIMISKAMING, etc.		UPPER HURONIAN *Unconformity* LOWER HURONIAN		
Great unconformity		*Great unconformity*		*Unconformity*		*Unconformity*		
(LAURENTIAN INTRUSIVES)		(Granites eruptive through the Keewatin)		Granite intrusives		LAURENTIAN (granite-gneiss, batholithic in Ontarian)	ARCHEAN	LAURENTIAN
						Irruptive contact		
LOGANIAN { Grenville (*Sedimentary*) Keewatin (*Igneous*) }		KEEWATIN, probably equivalent of Grenville		Keewatin group		ONTARIAN { Keewatin Couchiching }		KEEWATIN

* *Ont. Bur. Mines*, XXII, Part II, Appendix.
† *Ibid.*, XXIII, Part I, p. 206.
‡ *Compte-Rendu, 12th Int. Geol. Congress*, p. 403.
§ *Ibid.*, p. 351.
‖ Pp. 42 and 43.

most fully developed, appear to have an approximate aggregate thickness of 29,000 feet.[1] Elsewhere these sediments are also known to have great thickness.

The term "Eparchean Interval," concerning which there has been so much discussion in the past, should now be discarded. The so-called Lower and Middle Huronian rocks of the classic area of Lake Huron lie above and not beneath it. There are now known to be two great "intervals" in the pre-Cambrian, that beneath our Animikean and that beneath our Timiskamian, and not merely one as has been assumed.

Moreover, a dual subdivision of the pre-Cambrian cannot be justified on the basis of differences in metamorphism or other characteristics of the rocks. Normally there is a gradual increase in degree of metamorphism from the youngest to the oldest. In the province of Ontario the Animikean rocks are only slightly metamorphosed or disturbed, while the Timiskamian are disturbed and frequently rendered schistose, and the Loganian are highly metamorphosed.

Sediments, Grenville, occur in great volume in the Loganian. Hence the "Archean" cannot be said to be essentially igneous, as assumed at one time. Moreover, ordinary stratigraphic methods, which have been cited as a test as to whether certain groups of rocks should be classed as Archean or otherwise, can be applied in studying the Grenville. This has been shown by the authors' work in southeastern Ontario.[2]

Briefly, it may be said that the discovery of the true relations of the Timiskamian destroys the argument for a dual classification of the pre-Cambrian.

COMMENTS ON TABLES I AND II

The name Keweenawan in the tables is employed in the sense in which it has been used by most writers on the pre-Cambrian of North America. Certain greenstones in the Lake Huron area were classed by Logan as intrusives of Huronian age. They have come to be known as the "Thessalon greenstones," after the name

[1] *Ont. Bur. Mines*, XXIII, 214.

[2] *Ibid.*, XXII, Part II, pp. 41 f.

of the near-by town, and during the last decade they have been frequently referred to in literature and classed as of Keewatin age. In the summer of 1914, however, the junior author was able to verify Logan's conclusion regarding their age, and to prove that they are in intrusive contact with the Animikean, as defined in Table I. Hence they are classed as Keweenawan.

The group name Animikean is explained in a preceding paragraph as well as in Table I. In addition to what has long been called the Animikie, it includes the usually flat-lying sediments of Lake Huron, which have been classed as Lower and Middle Huronian by various authors. There is an unconformity within these sediments but we do not think it is of the magnitude that it has been inferred to be. It is to be described in a forthcoming paper.[1] We look on this unconformity as being similar to those which have been observed in the Cobalt series. As already said, the flat-lying sediments of Lake Huron, the Ramsay Lake series of Sudbury, and the Cobalt series, all being of post-Algoman age and no discordance having been proved to exist between them and the Animikie, are classed by us with the latter rocks under the name Animikean. It may be added that the quartzites and jasper conglomerates of the upper part of the Lake Huron series resemble rather closely the upper part of the Cobalt series. As shown in Table II, Collins applies the local name Whitewater series to the sedimentary rocks of the Sudbury Nickel Basin. The name was given, however, by Nickles and Foerste in 1905 to a part of the Richmond.[2] Hence, it should not be employed for the Sudbury rocks. The name Fabre, which has been used by certain authors as synonymous with Sudbury or Timiskaming, should also be discarded, since the rocks to which the name was applied are merely part of the Cobalt series.[3]

The name Algoman was introduced by A. C. Lawson in 1913.[4] The authors, having found, in the areas examined by them, that granite and gneiss of post-Timiskamian and pre-Animikean age are of wide extent, had employed local names to designate these rocks. At Cobalt the granite was called Lorrain and in southeastern

[1] *Ont. Bur. Mines*, XXIV.

[2] *U.S. Geol. Survey, Professional Paper* 71, p. 172.

[3] *Ont. Bur. Mines*, XIX, Part II, p. 62.

[4] *Compte-Rendu, 12th Int. Geol. Congress*, pp. 359–61.

Ontario, Moira, while other writers had given the name Killarney to the granite that occurs along the north shore of Lake Huron. Algoman now being preferred, although not having priority, the other names may be discarded or used locally.

Basic rocks, of pre-Algoman and post-Timiskamian age, described in a preceding paragraph, are widespread, but no age name is given to them in the tables.

The age relations of the Timiskamian rocks have already been described, and reasons have been given for preferring the use of this name to the other that has been proposed for them.

On a preceding page reasons have been advanced for applying the name Laurentian to the pre-Timiskamian granites and gneisses and not to those of post-Timiskamian age.

The authors agree with Lawson that the Keewatin and the Grenville, or Couchiching, should be grouped together. Ontarian, however, should not be employed as the name of a pre-Cambrian group, since the New York State geologists have long used the term Ontaric (or Ontario) as synonymous with Siluric. Recognizing the objection to Lawson's use of the term, we have substituted for it the name Loganian, as shown in Table II.[1]

According to the views of the authors, the relations between the Keewatin and the Grenville are as follows: The Keewatin rocks represent, for the most part, submarine lava flows. On the surface of these flows are deposited the Grenville sediments. Moreover, it is believed that, while the major part of the Grenville is later than the major part of the Keewatin, a minor part of one group is contemporaneous with a minor part of the other. Hence the two groups, volcanic and sedimentary respectively, are placed in the authors' classification under one age term, the Loganian. Lawson's Couchiching, it seems to the authors, should be correlated with the base of the Grenville, and, since Grenville is the older term, we prefer to discard Couchiching or use it only locally.

That part of the Loganian known as the Grenville is found in much greater volume in southeastern Ontario and the adjacent part of Quebec than in the northern and northwestern parts of the

[1] *Ont. Bur. Mines*, XXII, Part II, Appendix.

province. This is apparently chiefly due to less erosion having taken place, especially prior to Timiskamian time, in the southeastern section than in the others. Over the northern and northwestern areas the Grenville was eroded until only comparatively small remnants in most localities were left. Later erosion, and intrusions, especially the Algoman, have also assisted in destroying the Grenville. However, the broader, more extensive areas of the series in the southeast can be connected, by tracing outliers from point to point, with the exposures in the north and northwest, where occur Timiskamian and Animikean rocks.

Coleman has described a comparatively large area of Grenville rocks, seven or eight miles southeast of the town of Sudbury, whose relations to the Timiskamian, his Sudbury series, he has determined.[1] The Grenville here consists of cyrstalline limestone, quartzite, and various schists and gneisses, while the Timiskamian contains quartzites, greywackes, and other sediments. The Grenville series is profoundly metamorphosed. The Timiskamian, on the other hand, while often schistose and resting in highly inclined positions, has suffered much less metamorphism than the Grenville. There can be no doubt that the two series near Sudbury are separated by a great discordance and that the Grenville is vastly older than the Timiskamian. Similar relationships are found in other localities in most of which the Grenville is represented by only very small outliers.

Near the eastern end of Lake Kipawa, Quebec, which lies not far distant from the foot of Lake Timiskaming, there is an area of Grenville, consisting of crystalline limestone and other sediments. These rocks, described by the senior author in 1901, have been much disturbed by an intrusion of granite, apparently of the same age as that of Lake Timiskaming, on the eroded surface of which lie the fragmental rocks of the Cobalt series, Animikean.[2]

It would thus seem that the authors are justified in correlating the rocks of southeastern Ontario with those to the north and northwest. They have no doubt as to the relations of the Grenville to the Keewatin which occurs in large volume in the southeastern part

[1] *Ont. Bur. Mines,* XXIII, 208–9.

[2] *Am. Geologist,* January, 1901; *Ont. Bur. Mines,* XIII, 102.

of the province as well as in the north and northwest. In the southeastern section there is also a series, the Hastings, that is believed to represent the Timiskamian of the north.[1]

METALLOGENETIC NOTES

During the last decade, owing to the great progress that has been made in the production of metals in Ontario, special facilities have been provided for the study of pre-Cambrian rocks. Our information has been much increased concerning the age relations not only of the rocks that represent various epochs of this great period but also of the ore deposits that are associated with them. Of no other part of the continent, or of the world, has the pre-Cambrian proved to be of greater economic interest. The province not only has the world's greatest deposits of nickel, among which have been developed mines that compare favorably in economic importance with those of any other metal found elsewhere, but the gold mines and the cobalt-silver areas are recognized as being among the greatest known.

From Table III it will be seen that there were at least four great metallogenetic epochs during the pre-Cambrian period in Ontario—Grenville, Algoman, Animikean, and Keweenawan. A fifth epoch of minor importance should probably be added to represent the ore bodies associated with the basic intrusives that preceded the intrusion of the Algoman granite and followed the deposition of the Timiskamian sediments. There is proof that many important ore deposits have been removed by erosion, and it seems not unlikely that the rocks of certain epochs, not now productive, contained deposits which have disappeared through the removal of vast thicknesses of material.

SEQUENCE OF INTRUSION AND METAL DEPOSITION

Table III brings out an interesting alternation of intrusion and sedimentation, and the importance of the igneous rocks in the formation of ore deposits. It will be seen that there are broadly five great epochs of igneous activity, basic and acid rocks alternating, viz.: (1) Keewatin, basic; (2) Laurentian, acidic; (3) pre-Algoman, basic; (4) Algoman, acidic; (5) Keweenawan, basic,

[1] *Ont. Bur. Mines,* XXII, Part II.

passing in places into a considerable volume of acidic varieties. The pre-Algoman basic rocks are of greater volume and wider extent than they are usually recognized to be, since they are frequently wrongly classed as Keewatin. These basic rocks are

TABLE III

PRE-CAMBRIAN EPOCHS OF ONTARIO AND THEIR METAL PRODUCTION

KEWEENAWAN	Epoch, following basic intrusions, of (*a*) silver, cobalt, nickel, and arsenic at Cobalt and elsewhere, (*b*) nickel and copper at Sudbury, and copper elsewhere. Certain gold deposits, not now productive, appear to belong to this epoch.
ANIMIKEAN	Epoch of deposition of "iron formation" as a chemical precipitate.
(ALGOMAN)	Epoch, following granite intrusions, of gold at Porcupine and at many other localities, and of auriferous mispickel. Deposits of galena, zinc blende, fluorite, and other minerals also appear to have been derived from the granites, but some of them were not formed till post pre-Cambrian time. Preceding the intrusion of the Algoman granites, basic intrusives, that appear to be of post-Timiskamian age, gave rise to nickel and titaniferous and non-titaniferous magnetite deposits and chromite.
TIMISKAMIAN	Epoch of minor deposition of "iron formation" as a chemical precipitate.
(LAURENTIAN)	Granite intrusions probably gave rise to ore deposits which have been removed by excessive erosion as is known to be the case with deposits of later origin.
LOGANIAN	
Grenville	Epoch of deposition of extensive "iron formation" as a chemical precipitate among other sediments.
Keewatin	Composed largely of basic volcanic rocks.

represented by the sudburite of the Sudbury area, by the lamprophyres of Cobalt and elsewhere, and apparently by the basic rocks of the townships of Dundonald, Reaume, and others where associated with them are nickeliferous pyrrhotite and chromite.

Owing to erosion, the sequence of metal deposition shown in Table III is doubtless incomplete.[1] Iron formation occurs in three

[1] Two or three hundred feet more of erosion would have left comparatively little, for example, of the Cobalt silver deposits, of which, it would seem, more has been eroded than has been mined, or even of the great Mesabi iron deposits of Minnesota.

epochs, the Loganian, Animikean, and Timiskamian, but is of economic importance only in the former two. Certain deposits of titaniferous and non-titaniferous magnetites, not now being worked, are associated with basic intrusives that appear to be of pre-Algoman age. Arsenic occurs in two epochs and has been produced in economic quantities from the rocks of both. In so far as is known, gold occurs in economic quantity only in the Algoman, although small quantities are obtained in refining the copper-nickel ores, and certain auriferous quartz deposits, not now productive, appear to be genetically connected with Keweenawan intrusives. Nickel, as has been shown in the preceding table, was deposited in economic quantities in two epochs. Cobalt, silver, and copper are produced only from deposits of Keweenawan age. Platinum, palladium, mercury, and other metals are found in small quantities with Keweenawan ores. Zinc and lead have been mined in the province, but the age relations of some of the deposits are in doubt.

RELATIVE ECONOMIC IMPORTANCE OF VARIOUS EPOCHS

In Ontario in 1913, the metal production from ore deposits of various pre-Cambrian epochs was as follows: Keweenawan (silver, copper, nickel, cobalt, arsenic), $36,559,688; Algoman (gold), $4,543,690; Keewatin-Grenville (iron and iron ore), $1,195,150. A comparatively small quantity of the nickel and copper credited to the Keweenawan should be assigned to deposits that are associated with basic eruptives of pre-Algoman, and probably post-Timiskamian age, but the value is not definitely known.

While at present in Ontario there is no production from deposits of Animikean or Timiskamian age, millions of tons of iron ore are mined from deposits of these epochs in the state of Michigan.

It is to be understood that the ages given for the deposits do not refer to secondary concentration, as, for instance, in the case of iron ores, but to the epoch in which the metals were first deposited.

In a paper entitled "Metallogenetic Epochs in the Pre-Cambrian of Ontario," presented before the Royal Society of Canada, May, 1915, the authors discuss more fully the age and genetic relations of the ore deposits of the province.

METAMORPHIC STUDIES[1]

CONVERGENCE TO MINERAL TYPE IN DYNAMIC METAMORPHISM

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The thesis of this paper is that the formation of slates, schists, and some gneisses, by rock flowage, requires both mineralogical and chemical changes, and that there is convergence, both chemically and mineralogically, toward a comparatively few platy or columnar minerals, tending to give these rocks their characteristic lamellar structure. If such convergence can be demonstrated, it may be used as a guiding principle in the study and interpretation of this phase of metamorphism. In this paper it is proposed to discuss anamorphism from this point of view.

Mineral changes in anamorphism by rock flowage.—It is generally recognized, and need not here be demonstrated, that anamorphism by rock flowage tends to produce platy and columnar minerals, like mica, chlorite, hornblende; that these increase at the expense of other minerals during the process, and that in the resulting slates, schists, or gneisses, they are the characteristic minerals which determine the lamination and cleavage of the rock and give the rock its name. Quartz and feldspar are also important minerals in the slates, schists, and gneisses, but these minerals are less important than the first-named group in producing the essential characteristics of the slates, schists, and gneisses, and they tend to diminish relatively in quantity in proportion as platy and columnar minerals increase. Mica, chlorite, hornblende, feldspar, and quartz make up all but an insignificant part of the mass of schists, slates, and gneisses. There are of course other minerals developed, such as garnet, staurolite, chloritoid, sillimanite, andalusite, etc., which in certain rocks may be conspicuous and

[1] See also C. K. Leith, "The Metamorphic Cycle," *Jour. Geol.*, XV (1907), 303–13; C. K. Leith and W. J. Mead, "Metamorphic Studies," *ibid.*, XX (1912), 353–61; C. K. Leith and W. J. Mead, *Textbook of Metamorphism* (in press, Henry Holt & Co.).

distinctive, and therefore are used in naming the rock, but in abundance they must be regarded as distinctly minor and accessory constituents as compared to the principal constituents before named.

In the progressive changes from a mud or clay to a shale and in turn to a slate or schist, there is an increase in the percentage by weight of mica and chlorite with relative decrease of other constituents. Muscovite or sericite makes up from a third to a half of mica slates, according to Dale,[1] and in a phyllite may be in even larger proportion, whereas in the original rock the percentage was but a small fraction of this amount. There is corresponding decrease in kaolin, feldspar, and certain ferromagnesian minerals.

Where a sandstone or quartzite becomes schistose there is an increase of mica, with corresponding decrease of other constituents. Certain specific instances of development of sericite schists in shear zones crossing the bedding of a quartzite have shown elimination of the greater part of the quartz of the original rock, due to solution, probably aided greatly by minute granulation.

A limestone or a marble may become schistose purely by process of granulation, which involves no essential development of new minerals. But more often there is a striking development of amphibole and other silicates with a corresponding diminution in calcite. Silicated marbles, resulting from metamorphism, are too well known to need further description.

A granite, by metamorphism during rock flowage, may become a mica schist or mica gneiss. Mica is increased at the expense of the feldspar and some of the ferromagnesian minerals.

A basic igneous rock, by dynamic metamorphism, may become a chlorite or hornblende schist or gneiss, the increase of chlorite or hornblende being counterbalanced by decrease in feldspar and pyroxene.

The average igneous rock, according to Clarke,[2] has about 60 per cent feldspar, 28 per cent ferromagnesian and accessory

[1] T. Nelson Dale, "Slates of the United States," *Bull. 586, U.S. Geol. Survey*, 1914, p. 20.

[2] F. W. Clarke, "The Data of Geochemistry," *Bull. 491* (2d ed.), *U.S. Geol. Survey*, 1911, p. 398.

minerals, and 12 per cent quartz. The schists or gneisses developed from these igneous rocks have a considerably smaller proportion of feldspars. The ferromagnesian minerals, instead of consisting dominantly of augite and biotite, with some muscovite, are now largely hornblende and muscovite, with subordinate quantities of biotite.

Whether the parent rock is igneous or sedimentary, or whatever its mineral content, the resulting schists and gneisses are characterized by hornblende, chlorite, and mica, which are developed to such an extent that the very nature of the original rock is often lost. If the mineralogical change were not extensive, the problem of origin of schists and gneisses would not be nearly so difficult as it is. The very existence of the problem testifies to the great mineralogical changes which have occurred. A sericite schist, for instance, may result from dynamic anamorphism of a slate, an igneous rock, or a quartzite. A hornblende schist may result from the anamorphism of a limestone or marble or may develop from a basic igneous rock, a graywacke or other basic sediment, or a slate.

A few distinctive minerals, such as mica, chlorite, and hornblende, are the characteristic resulting products of anamorphism by rock flowage of a considerable variety of parent rocks. In this sense, then, there is a convergence toward a certain mineralogic type.

Chemical changes during anamorphism by rock flowage.—When the chemical changes in anamorphism by rock flowage are considered, there is less general acceptance of the fact that actual changes in composition take place and that these changes are in the direction of producing the composition of the characteristic minerals in the end-products. Probably it is the general view that the conditions of rock flowage and anamorphism are not usually favorable to extensive transfers of materials involved in change of composition. It is often true that there are no important changes in percentages of chemical constituents during the process of rock flowage and anamorphism, the mineralogical change having been accomplished through recrystallization of substances present. This is especially true when the original rocks are well adapted, in

their composition, like shales, to the production of the necessary schist-making minerals. But even in these cases the elimination of water, carbon dioxide, and oxygen is recognized. We think it can be likewise inferred, in cases where the composition of the parent rock differs widely from the composition of the hornblende, mica, or chlorite, that important chemical changes take place, principally by the elimination of the substances present in excess of these requirements, but possibly also in some cases by addition of substances from without, and that there is therefore a chemical as well as mineralogical convergence toward mineral types.

Illustrative of such changes is the diminution of carbon dioxide, and perhaps lime, when a marble becomes silicated. In extreme cases an amphibolite has been described as the end-result of the alteration of marble. This obviously means a considerable change in chemical composition. For the purposes of our argument it is immaterial whether this change consists dominantly of elimination or of substitution of constituents from without along igneous contacts. Rock flowage and anamorphism are often accomplished under conditions which make it impossible to determine the relative importance of the purely mechanical and of the igneous influence respectively in their production. The point we would emphasize is that the characteristic product like amphibolite or silicated marble resulting from rock flowage or anamorphism differs distinctly in composition from the primary limestones or marble.

The development of talc schist from the anamorphism of a dolomite indicates an important change of composition, like that in the Menominee district of Michigan.[1]

A quartzite often becomes sheared and transformed into a sericite schist. Where shearing follows impure phases of the quartzite parallel to the bedding, it is difficult to prove any change in composition, but where the sericite schist develops in shear zones, crossing massive beds of quartzite, and samples can be collected showing complete gradation of the massive quartzite into the sericite schist, there is conclusive evidence of the elimination of quartz. The authors have carefully sampled two good cases of

[1] W. S. Bayley, "Menominee Iron-bearing District of Michigan," *Mon. 46, U.S. Geol. Survey*, 1904.

this kind in the Marquette district of Michigan. Another case is worked out in similar fashion by J. H. Warner on the Waterloo quartzite of Wisconsin, where shear zones cross the bedding. Analyses of these rocks,[1] show that the larger part of the free quartz has been eliminated, presumably through solution aided by granulation.

Where shales are traced into slates or schists the change in composition is less obvious, owing to the fact that the shale originally possesses pretty nearly the composition required by the end-products, but even here, important changes in composition are noted, like the elimination of CO_2, water, and oxygen, and, in some cases, the increase of potash.

A greenstone in the Marquette district of Michigan has been traced inch by inch into a sericite-chlorite schist and sampled on a large scale, the resulting analyses showing elimination of silica, an increase of alumina and ferric oxide, potash, and a considerable decrease in lime. Similar changes are described and analyzed by George H. Williams.[2]

It is not easy to find in the field satisfactory gradations from massive to schistose rocks where one can be certain that the mass was throughout of uniform composition. It is still more rare that these gradations have been thoroughly sampled so that the analyses furnish an adequate basis for comparison. It is significant, however, that where this has been done—and we have searched the literature carefully—important changes are to be noted.

Fortunately we are not obliged in our conclusion to rely entirely on the few well-sampled gradations from primary to schistose rock. A study of the actual compositions of the principal groups of schists brings out the fact that these vary in certain essential respects from those of the primary rocks from which they are supposed to have been derived. In another discussion[3] of the use of chemical criteria in the identification of schists and gneisses,

[1] C. K. Leith and W. J. Mead, *Textbook of Metamorphism* (in press, Henry Holt & Co.

[2] G. H. Williams, "The Greenstone Schist Areas of the Menominee and Marquette Regions of Michigan; A Contribution to the Subject of Dynamic Metamorphism in Eruptive Rocks," *Bull. 62, U.S. Geol. Survey*, 1890.

[3] *Textbook of Metamorphism*, cited above.

analyses of common schists have been platted graphically in order that they may be compared with both known and possible parent rocks. An examination of these plates brings out the fact that the composition of the schist tends to approach the distinctive chemical characteristics of the dominant platy or columnar mineral in the schist. This is especially well shown in the sericite schist, the composition of which lies between that of the primary rock and that of the mineral sericite. It is indicated, not only by the position of the analyses on the diagram, but by the shapes of the flags showing the relative amounts of constituents. If we were to include in these plates only those schists in which the processes of anamorphism and rock flowage have gone to an extreme, the tendency would be still more obvious, because rocks are included in these diagrams, described as hornblende, chlorite, or mica schist, which have but a small proportion of these minerals.

Our inference from the available facts is that, while recrystallization of substances present has of course played an important part in the production of schists, for some rocks important changes in composition have also occurred; that these changes in composition have been toward the composition of the characteristic end-products—mica, hornblende, or chlorite; that these changes are known both in sedimentary and in igneous rocks, of both acid and basic composition, and that the changes have been sufficiently important to make it impossible, along with other reasons, to use chemical composition as a conclusive criterion for the identification of origin of schists and gneisses.

Significance of convergence.—If this idea of convergence be correct, our attention is directed to the physical and chemical characteristics of a few minerals like mica, chlorite, and hornblende as important factors in anamorphism by rock flowage. Obviously they are adapted to the conditions of rock flowage; otherwise they would not develop at the expense of other minerals. It is not so clear whether they are adapted by their crystal habits, by their cleavages, by their composition, or by a combination of these characters. The fact that they are always arranged according to their dimensions, their greatest and least mean being respectively parallel, and the fact that their dimensions in the schist are clearly those

determined by their habit of crystal growth, rather than by any growth or breaking giving other shapes, direct our attention to the crystal habit of these minerals as essential reasons for their development. Whether processes of rock flowage have been weak or intense, the resulting minerals maintain their habit and dimensional characteristics.

If convergence toward certain characteristic minerals is established in rock flowage, the question naturally arises whether these end-products can be regarded purely as results of the great variety of processes in anamorphism by rock flowage, or whether the individual characters of the resulting minerals have exerted a certain directive and controlling influence in converging the various lines of anamorphism and rock flowage toward themselves, and more particularly whether the crystal habit of these minerals has exerted this influence. From this point of view it seems to be something more than a coincidence that, under such a variety of conditions and with such variety of available materials, that which we call crystallizing force has been able to exert itself to the extent of causing important mineralogical and chemical changes toward a limited number of mineral forms. It has drawn to itself the materials needed, eliminated those not needed, and has developed crystals of uniform habit. Under given environment it has had the capacity to organize the substances in a fashion best adapted to environment in much the same way that organisms have been supposed to adapt themselves to environment. The problem before us may be similar to the biologic question whether organisms are distinctly the results of physical and chemical environment or whether that mysterious force we call life on occasion rises superior to environment and to some extent modifies and controls external conditions. It is sometimes said that man is the multiplier and environment the multiplicand in the product determining history. The attempt to apply the same reasoning to the development of certain characteristic minerals in schists suggests the crystallizing power of these crystal individuals as the multiplier and environment as the multiplicand in the product representing rock flowage.

It is of interest also to note that the convergence here argued is toward a group mineralogically and chemically different from

primary igneous rocks, from which all metamorphism presumably starts, and is essentially a by-product of the metamorphic cycle. The difference between schists, gneisses, and slates, on the one hand, and igneous rocks on the other, represents a gap which has not been closed by the metamorphic cycle. The schists, gneisses, and slates represent types which are clearly nearer the composition of the original igneous rock than many of the rocks, particularly the sediments, which have been anamorphosed and recrystallized and have undergone rock flowage.

This convergence toward igneous rock composition often reaches a point where chemical criteria are not sufficient to determine whether or not the primary rock was igneous or sedimentary, the resulting composition being so nearly that of an igneous rock. If, along igneous contacts, a considerable amount of material has been completely fused, the anamorphic product may take on the character of an igneous rock and in such places the cycle is closed. But just as sandstones or limestones may be regarded as products of the cycle, some of which never go back to their primary condition of igneous rocks, so the schists and gneisses are by-products, which do not go back to the condition of igneous rocks under ordinary conditions of anamorphism. Whether the cycle in its larger aspects is completely closed, so far as the great mass of products is concerned, is a very doubtful question which has been discussed in a previous paper.[1]

[1] *Jour. Geol.*, XX (1912), 353–61.

THE AGE OF THE CRETACEOUS FLORA OF SOUTHERN NEW YORK AND NEW ENGLAND

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The age of the plant-containing Upper Cretaceous which outcrops, or is found in morainal deposits, on Staten Island, along the north shore of Long Island, and on Block Island and Martha's Vineyard has never been exactly determined, although their obvious general relations may be indicated by the fact that as early as 1838 Mather[1] suggested their equivalency with the New Jersey beds around Raritan Bay.

Only one or two points in the history of their study need be mentioned. Newberry considered them the equivalent of the Amboy clays which as then developed paleobotanically were almost entirely Raritan. Ward[2] named them the Island series and considered them uppermost Potomac but post-Raritan, the latter being erroneously regarded as of late Lower Cretaceous age.[3] White,[4] who was the first to prove the existence of Upper Cretaceous on Martha's Vineyard is quoted by Veatch (1906) as holding that the plant beds on that island are about equivalent to those at Cliffwood, New Jersey, an undoubtedly correct interpretation.

It was not until the Magothy formation was delimited from the underlying Raritan formation and the overlying Matawan formation by Clark[5] in 1904 and its flora described by the writer[6] that adequate paleobotanical data were available for exact comparisons with the more or less isolated and disturbed outcrops to the eastward.

[1] W. W. Mather, *Rept. First Dist. N.Y.* (1838), pp. 136–37.

[2] Lester F. Ward, *Fifteenth Ann. Rept. U.S. Geol. Surv.* (1895), p. 335.

[3] It is not worth while to consider the claim of O. C. Marsh that these beds represent Jurassic sediments.

[4] D. White, *Am. Jour. Sci.* (III), XXXIX (1890), 93–101.

[5] W. B. Clark, *Am. Jour. Sci.* (IV), XVIII (1904), 435–40.

[6] E. W. Berry, *Ann. Rept. State Geol. N.J. for 1905*, pp. 135–56.

In the last few years the geology of Long Island has been studied by Veatch,[1] Fuller,[2] and others, and the southward extensions of the Raritan and Magothy floras across New Jersey and in Delaware and Maryland have been investigated by the writer.[3] From the strike of the Cretaceous beds across Long Island as determined by the authors above mentioned; by a study of their lithology and genesis; and by an analysis of the insular floras, for a knowledge of which we are indebted almost entirely to the long-continued labors of Dr. Arthur Hollick,[4] it is now possible to determine with certainty the age of a number of the localities where Cretaceous plants have been found and to show with a fair degree of probability the age of the original deposits from which the more outlying morainal-materials-carrying plants have been derived.

Without burdening the reader with the steps in this analysis, the conclusions at which the writer has arrived will be given, and a list of the Raritan and Magothy floras will be appended, so that the following statements can be verified by anyone that cares to take the trouble to do so.

The beds on Staten Island offer little difficulty, since they are practically continuous with those of New Jersey; in fact Hollick has indicated in several publications that the celebrated Kreischerville outcrops are of Raritan age. From clays in place plants have been collected at Green Ridge and around Kreischerville. From morainal material Upper Cretaceous plants are recorded from Tottenville, Richmond Valley, Princess Bay, and Arrochar. These also are probably all of Raritan age.

Along the north shore of Long Island fossil plants have been collected from clays in place at Glen Cove, Cold Spring, and Northport. From the geographical position of these outcrops, their distance above bedrock, and as is conclusively shown by

[1] A. C. Veatch, *Professional Paper No. 44, U.S. Geol. Surv.*, 1906.

[2] M. L. Fuller, *Professional Paper No. 82, U.S. Geol. Surv.*, 1914.

[3] E. W. Berry, Contributions to the Mesozoic Flora of the Atlantic Coastal Plains: I, *Bull. Torrey Bot. Club*, XXXIII (1906), 163–82, pls. 7–9; IV, *Ibid.*, XXXVII (1910), 19–29, pl. 8; VII, *Ibid.*, XXXVIII (1911), 399–424, pls. 18, 19; X, *Ibid.*, XLI (1914), 295–300.

[4] Nearly all the literature is summarized by him in *Mon. U.S. Geol. Surv.* (1906), L, since which date only a few minor papers have been published.

their extensive flora, they are unquestionably of Magothy age. Fossil plants have been found in morainal materials at a large number of localities on Long Island. These are Brooklyn, Elm Point (Great Neck), Mott Point (Manhasset Neck), Glen Cove, Sea Cliff, Dosoris Island, Lloyd Neck, Eaton's Neck, Roslyn, Oak Neck, and Montauk Point.

While it is quite possible that some of these materials may have been derived from Raritan outcrops once present to the northward, the flora does not indicate Raritan affinities. Of the 100 species listed from Long Island in *Monograph 50* (*op. cit.*), only 33 have been found in the New Jersey Raritan, which is about the same proportion that survives from the Raritan into the overlying Magothy formation in the New Jersey area. There is then no paleobotanical evidence of the existence of the Raritan east of Brooklyn.

The probable genesis of the Raritan has an important bearing on this question. The Raritan formation in the area of its type development (i.e., around Raritan Bay, New Jersey, the so-called Amboy district) consists of a lower, middle, and upper horizon with thick, argillaceous beds, separated by well-defined sand beds from 35 to 50 feet in thickness. The Raritan fire and potter's clay at the base is about 35 feet thick. The middle or Woodbridge clays range from 30 to 60 feet in thickness, while the upper clays (South Amboy fireclay and stoneware clay) are about 50 feet in thickness. These all carry fossil plants. In discussing this flora the writer has demonstrated[1] that it shows a remarkable cleavage into an older flora (Lower and Middle Raritan) and a younger flora (Upper Raritan), from the latter of which a large number of the forms common to the Magothy take their origin. In tracing the Raritan southwestward across New Jersey, Delaware, and Maryland to the Potomac River valley it has been found impossible to recognize the well-marked divisions described in the Amboy district. The massive clays of the latter are replaced within a dozen miles by prevailingly arenaceous deposits, and with the exception of certain well records, e.g., Fort Dupont, Delaware,

[1] E. W. Berry, "The Flora of the Raritan Formation," *Bull. 3, Geol. Surv. of New Jersey*, 1911.

which are open to various interpretations, the formation is everywhere much thinner. This is capable of more than one explanation, but from a consideration of all the facts it seems probable that at least the older Raritan, that is to say the Lower and Middle Raritan of the Amboy district from the base to the top of the Woodbridge clays, the upper surface of which shows considerable erosion, is a purely local deposit continental or estuarine in character, and that the southwesterly extension of the Raritan represents the Upper Raritan. In endeavoring to trace the Raritan eastward beyond Staten Island one searches in vain in the very numerous well records (Veatch listed over 900 in 1906) for the representatives of the Raritan, Woodbridge, and Amboy clays, the basal beds of the coastal plain in the eastern area being prevailingly sandy with irregular clay lenses of no great extent. This together with the paleobotanical evidence leads to the conclusion that the Lower and Middle Raritan were never present east of Brooklyn, and that while the Upper Raritan may have extended eastward as it did to to the southwestward there is at present no evidence of such an extension. If the Long Island Cretaceous were Upper Raritan, the described floras are extensive enough to determine this point.

Obviously, if the Raritan failed to extend along the north shore of Long Island it was never represented on Martha's Vineyard, Block Island, or the Elizabeth Islands, and this is strikingly confirmed by the lithologic character of the Upper Cretaceous in the latter region and by the extensive flora, as has been recognized to a more or less degree by White, Ward, and Hollick, who have studied this flora. Of the 126 recorded species from Martha's Vineyard and Block Island only 36 are found in the Raritan. The number common to the Magothy is but 37, but these include a large number of forms that are distinctive horizon markers and range to higher levels in our southern states.

According to the age determinations as outlined above, the Raritan and Magothy floras are segregated in the following lists. These show that the Raritan flora comprises 224 and the Magothy flora 289 species (so called); 61 per cent of the Raritan flora does not extend into the Magothy and 70 per cent of the Magothy flora does not occur in the Raritan. A mere statistical method does

not, however, sufficiently emphasize their contrasts. The Raritan contains a number of types that survived from the Lower Cretaceous, while the Magothy introduces many modern types in genera, among which the following may be mentioned:

Amelanchier	Doryanthites	Lycopodium	Paliurus	Planera
Araucaria	Elaeodendron	Marsilea	Panax	Rhamnus
Bumelia	Guatteria	Nectandra	Periploca	Rhus
Coccolobites	Gyminda	Nelumbo	Persea	Sabalites
Crataegus	Hedera	Ocotea	Picea	Viburnum
Dalbergia	Illicium	Onoclea	Pistia	Zizyphus

Only three of these genera are sparingly represented in the Raritan. It would hardly be worth while to follow the analysis farther in the present connection, as sufficient facts have already been given to establish the probability of the questions considered. Both Veatch and Fuller (*op. cit.*) consider that some of the north-shore outcrops represent the Raritan. Veatch, in discussing the underground waters of Long Island in 1906 (*op. cit.*) differentiated a horizon, the Lloyd sand, whose top is from 150 to 200 feet above bedrock. This sand, which may or may not be a continuous stratum, he believes he has identified in a considerable number of wells along both the north and south shores of Long Island and also in various records of wells in eastern New Jersey. He places it in the Raritan, but if it is really a continuous bed at the same stratigraphic level it probably represents the so-called laminated sands[1] formerly considered a part of the Raritan but shown by the writer to be lower Magothy[2].

Here follows lists of the plants identified from the Raritan and Magothy formations throughout their extent, which in the case of the latter is from Martha's Vineyard to the Potomac River valley, a distance along the strike of something like 450 miles (750 kilometers).

List of Plants Recorded from the Raritan Formation

Acer amboyense Newberry
Acer minutum Hollick
Acerates amboyense Berry
Andromeda grandifolia Berry
Andromeda Cookii Berry
Andromeda novae-caesareae Hollick
Andromeda Parlatorii Heer
Andromeda tenuinervis Lesquereux

[1] *Final Rept. State Geologist, N.J.*, VI (1904), 168.

[2] *Ann. Rept. State Geol. N.J., for 1905*, p. 136.

Androvettia statenensis Hollick and Jeffrey
Anomaspis hispida Hollick and Jeffrey
Anomaspis tuberculata Hollick and Jeffrey
Aralia formosa Heer
Aralia groenlandica Heer
Aralia Newberryi Berry
Aralia patens Hollick
Aralia quinquepartita Lesquereux
Aralia rotundiloba Newberry
Aralia washingtoniana Berry
Aralia Wellingtoniana Lesquereux
Araliopsis breviloba Berry
Araliopsis cretacea (Newberry) Berry
Araliopsis cretacea dentata (Lesquereux) Berry
Araliopsis salisbureaefolia (Lesquereux) Berry
Araucariopitys americana Jeffrey
Araucarioxylon noveboracense Hollick and Jeffrey
Aspidiophyllum trilobatum Lesquereux
Asplenium Dicksonianum Heer
Asplenium Foersteri Debey and Ettingshausen
Asplenium jerseyensis Berry
Asplenium raritanensis Berry
Baiera incurvata Heer
Bauhinia cretacea Newberry
Bauhinia gigantea Newberry
Brachyoxylon notabile Hollick and Jeffrey
Brachyphyllum macrocarpum Newberry
Caesalpinia Cookiana Hollick
Caesalpinia raritanensis Berry
Calycites diospyriformis Newberry
Calycites parvus Newberry
Carpolithus euomymoides Hollick
Carpolithus floribundus Newberry
Carpolithus hirsutus Newberry
Carpolithus ovaeformis Newberry
Carpolithus pruniformis Newberry
Carpolithus vaccinioides Hollick
Carpolithus woodbridgensis Newberry
Celastrus arctica Heer
Celastrophyllum Brittonianum Hollick
Celastrophyllum crenatum Heer
Celastrophyllum cretaceum Lesquereux
Celastrophyllum decurrens Lesquereux
Celastrophyllum grandifolium Newberry
Celastrophyllum minus Hollick
Celastrophyllum Newberryanum Hollick
Celastrophyllum spatulatum Newberry
Celastrophyllum undulatum Newberry
Chondrites flexuosus Newberry
Chondrophyllum obovatum Newberry
Chondrophyllum orbiculatum Heer
Chondrophyllum reticulatum Hollick
Cinnamomum Newberryi Berry
Cissites formosus Heer
Cissites Newberryi Berry
Citrophyllum aligerum (Lesquereux) Berry
Cladophlebis socialis (Heer) Berry
Colutea primordialis Heer
Comptonia microphylla (Heer) Berry
Cordia apiculata (Newberry) Berry
Cornophyllum vetustum Newberry
Cycadinocarpus circularis Newberry
Cyparissidium gracile Heer?
Czekanowskia capillaris Newberry
Dactyolepis cryptomerioides Hollick and Jeffrey
Dammara borealis Heer
Dammara minor Hollick
Dalbergia apiculata Newberry
Dalbergia hyperborea Heer?
Dewalquea groenlandica Heer
Dewalquea insignis Hosius and v.d. Marck
Dewalquea trifoliata Newberry
Dicksonia groenlandica Heer
Diospyros apiculata Lesquereux?
Diospyros primaeva Heer
Diospyros amboyensis Berry
Diospyros vera Berry
Eucalyptus angusta Velenovsky
Eucalyptus attenuata Newberry
Eucalyptus Geinitzi Heer
Eucalpytus linearifolia Berry
Eucalyptus parvifolia Newberry
Eugeinitzia proxima Hollick and Jeffrey
Ficus daphnogenoides (Heer) Berry
Ficus Krausiana Heer
Ficus myricoides Hollick

Ficus ovatifolia Berry
Ficus Woolsoni Newberry
Fontainea grandifolia Newberry
Frenelopsis Hoheneggeri (Ett.) Schenk
Geinitzia formosa Heer
Geinitzia Reichenbachi Hollick and Jeffrey
Gleichenia Giesekiana Heer
Gleichenia micromera Heer
Gleichenia Zippei Heer
Hedera obliqua Newberry
Hedera primordialis Saporta
Hymenaea dakotana Lesquereux
Ilex amboyensis Berry
Ilex elongata Newberry
Juglans arctica Heer
Juglans crassipes Heer
Juniperus hypnoides Heer
Kalmia Brittoniana Hollick
Laurophyllum angustifolium Newberry
Laurophyllum elegans Hollick
Laurophyllum lancealatum Newberry
Laurophyllum minus Newberry
Laurophyllum nervillosum Hollick
Laurus nebrascensis Hollick
Laurus Hollae Heer?
Laurus plutonia Heer
Leguminosites atanensis Heer
Leguminosites coronilloides Heer
Leguminosites omphalobioides Lesquereux
Leguminosites raritanensis Berry
Liriodendron oblongifolium Newberry
Liriodendron primaevum Newberry
Liriodendron quercifolium Newberry
Liriodendropsis angustifolia Newberry
Liriodendropsis retusa (Heer) Newberry
Liriodendropsis simplex Newberry
Magnolia alternans Heer
Magnolia Hollicki Berry
Magnolia Isbergiana Heer
Magnolia Boulayana Lesquereux
Magnolia Lacoeana Lesquereux
Magnolia longipes Hollick
Magnolia Newberryi Berry
Magnolia speciosa Heer
Magnolia woodbridgensis Hollick
Majanthemophyllum pusillum Heer
Menispermites borealis Heer
Menispermites Wardianus Hollick
Microzamia gibba (Reuss) Corda
Moriconia cyclotoxon Debey and Ettingshausen
Myrica acuta Hollick
Myrica cinnamomifolia Newberry
Myrica Davisii Hollick
Myrica emarginata Heer
Myrica fenestrata Newberry
Myrica Hollicki Ward
Myrica Newberryana Hollick
Myrica raritanensis Hollick
Myrsine borealis Heer
Myrsine Gaudini (Lesquereux) Berry
Myrsine oblongata Hollick
Newberryana rigida (Newberry) Berry
Paliurus affinis Heer?
Passiflora antiqua Newberry
Phegopteris Grothiana Heer
Phaseolites elegans Hollick
Phyllites poinsettioides Hollick
Phyllites trapaformis Berry
Pinus granulata (Heer) Stopes
Pinus quinquefolia Hollick and Jeffrey
Pinus raritanensis Berry
Pinus tetraphylla Jeffrey
Pinus triphylla Hollick and Jeffrey
Pistacia aquehongensis Hollick
Pitoxylon statenense Hollick and Jeffrey
Planera Knowltoniana Hollick
Platanus aquehongensis Hollick
Platanus Heerii Lesquereux
Persoonia Lesquereuxii Knowlton
Persoonia spatulata Hollick
Phaseolites manhassettensis Hollick
Pityoidelepsis statensis Hollick and Jeffrey
Podozamites acuminatus Hollick
Podozamites Knowltoni Berry
Podozamites lanceolatus (L. and H.) F. Braun
Podozamites marginatus Heer
Populus harkeriana Lesquereux
Populus orbicularis (Newberry) Berry
Prepinus statenensis Jeffrey
Protodammara speciosa Hollick and Jeffrey

Protophyllocladus subintegrifolius (Lesq.) Berry
Protophyllum multinerve Lesquereux
Prunus ? acutifolia Newberry
Pseudogeinitzia sequoiiformis Hollick and Jeffrey
Pterospermites modestus Lesquereux
Pterospermites obovatus (Newberry) Berry
Quercus ? novae-caesareae Hollick ?
Quercus raritanensis Berry
Raritania gracilis (Newberry) Hollick and Jeffrey
Rhamnites minor Hollick
Salix flexuosa Newberry
Salix inaequalis Newberry
Salix Lesquereuxii Berry
Salix Newberryana Hollick
Salix pseudo-Hayei Berry
Salix raritanensis Berry
Sapindus Morrisoni Heer
Sassafras acutilobum Lesquereux
Sassafras hastatum Newberry
Sassafras progenitor Hollick
? Sequoia concinna Heer
Sequoia heterophylla Velenovsky
Sequoia Reichenbachi (Geinitz) Heer
Smilax raritanensis Berry
Sphaerites raritanensis Berry
Sphenaspis statenensis Hollick and Jeffrey
Strobilites Davisii Hollick and Jeffrey
Strobilites microsporophorus Hollick and Jeffrey
Tricalycites major Hollick
Tricalycites papyraceus Newberry
Tricarpellites striatus Newberry
Thuya cretacea (Heer) Newberry
Thuyites Meriana Heer
Viburnum integrifolium Newberry
Widdringtonites Reichii (Ettingshausen) Heer
Widdringtonites subtilis Heer
Williamsonia problematica (Newberry) Ward
Williamsonia Reesii Hollick
Williamsonia Smockii Newberry

List of Plants Recorded from the Magothy Formation

Acer paucidentatum Hollick
Amelanchier Whitei Hollick
Andromeda Cookii Berry
Andromeda grandifolia Berry
Andromeda novae-caesareae Hollick
Andromeda Parlatorii Heer
Aralia Brittoniana Berry
Aralia coriacea Velenovsky
Aralia Towneri Lesquereux
Aralia groenlandica Heer
Aralia mattewanensis Berry
Aralia nassauensis Hollick
Aralia Newberryi Berry
Aralia Ravniana Heer
Araucaria bladenensis Berry
Araucaria marylandica Berry
Araucarites ovatus Hollick
Araucarites Zeilleri Berry
Arisaema cretaceum Lesquereux
Arisaema (?) mattewanense Hollick
Asplenium cecilensis Berry
Algites americana Berry
Baiera grandis Heer ?
Banksia pusilla Velenovsky ?
Banksites Saportanus Velenovsky
Bauhinia marylandica Berry
Betulites populifolius Lesquereux
Brachyphyllum macrocarpum Newberry
Brachyphyllum macrocarpum formosum Berry
Bumelia praenuntia Berry
Calycites alatus Hollick
Calycites obovatus Hollick
Carix Clarkii Berry
Carpites liriophylli Lesquereux
Carpites minutulus Lesquereux
Carpolithus cliffwoodensis Berry
Carpolithus drupaeformis Berry
Carpolithus floribundus Newberry
Carpolithus hirsutus Newberry
Carpolithus juglandiformis Berry
Carpolithus mattewanensis Berry
Carpolithus ostryaeformis Berry
Carpolithus septloculus Berry

Cassia insularis Hollick
Ceanothus constrictus Hollick
Celastrus arctica Heer
Celastrophyllum crassipes Lesquereux?
Celastrophyllum crenatum Heer?
Celastrophyllum elegans Berry
Celastrophyllum grandifolium Newberry?
Celastrophyllum Newberryanum Hollick
Celastrophyllum undulatum Newberry?
Chondrites flexuosus Newberry?
Cinnamomum crassipetiolatum Hollick
Cinnamomum Heeri Lesquereux
Cinnamomum membranaceum (Lesquereux) Hollick
Cinnamomum Newberryi Berry
Cissites formosus magothiensis Berry
Cissites Newberryi Berry
Citrophyllum aligerum (Lesquereux) Berry
Coccolobites cretaceus Berry
Cocculus cinnamomeus Velenovsky
Cocculus imperfectus Hollick
Cocculus inquirendus Hollick
Cocculus minutus Hollick
Colutea obovata Berry
Confervites dubius Berry
Cordia apiculata (Newberry) Berry
Cornus cecilensis Berry
Cornus Forchhammeri Heer
Crataegus monmouthensis Berry
Credneria macrophylla Heer
Crotonophyllum cretaceum Velenovsky
Cunninghamites elegans (Corda) Endl.
Cunninghamites squamosus Heer
Cupressinoxylon Bibbinsi Knowlton
Czekanowskia dichtoma Heer?
Dalbergia irregularis Hollick
Dalbergia minor Hollick
Dalbergia severnensis Berry
Dammara borealis Heer
Dammara cliffwoodensis Hollick
Dammara minor Hollick
Dammara northportensis Hollick
Dewalquea groenlandica Heer
Diospyros apiculata Lesquereux?
Diospyros primaeva Heer
Diospyros prodromus Heer?
Diospyros provecta Velenovsky
Diospyros pseudoanceps Lesquereux
Diospyros rotundifolia Lesquereux
Doryanthites cretacea Berry
Dryandroides quiercinea Velenovsky
Elaeodendron marylandicum Berry
Elaeodendron strictum Hollick
Embothriopsis presagita Hollick
Eucalyptus (?) attenuata Newberry
Eucalyptus Geinitzi Heer
Eucalyptus Geinitzi propinqua Hollick
Eucalyptus latifolia Hollick
Eucalyptus linearifolia Berry
Eucalpytus Schubleri (Heer) Hollick?
Eucalyptus Wardiana Berry
Ficus atavina Heer
Ficus cecilensis Berry
Ficus crassipes Heer
Ficus daphnogenoides (Heer) Berry
Ficus Krausiana Heer
Ficus Krausiana subsimilis Hollick
Ficus myricoides Hollick
Ficus reticulata (Lesquereux) Knowlton
Ficus sapindifolia Hollick
Ficus Willisiana Hollick
Ficus Woolsoni Newberry
Frenelopsis Hoheneggeri (Ettingshausen) Schenk?
Geinitzia formosa Heer
Gleichenia delawarensis Berry
Gleichenia gracilis Heer?
Gleichenia protogaea Debey and Ettingshausen
Gleichenia Saundersii Berry
Gleichenia Zippei (Corda) Heer
Guatteria cretacea Hollick
Gyminda primordialis Hollick
Hedera cecilensis Berry
Hedera cretacea Lesquereux
Hedera simplex Hollick
Heterofilicites anceps Berry
Hymenaea dakotana Lesquereux
Hymenaea primigenia Saporta
Ilex papillosa Lesquereux
Ilex severnensis Berry
Ilex strangulata Lesquereux
Illicium deletoides Berry
Juglans arctica Heer
Juglans crassipes Heer

Juglans elongata Hollick
Juniperus hypnoides Heer
Laurophyllum angustifolium Newberry
Laurophyllum elegans Hollick
Laurophyllum lanceolatum Newberry
Laurophyllum ocoteaeoides Hollick
Laurus antecedens Lesquereux
Laurus atanensis Berry
Laurus Hollae Heer
Laurus Hollickii Berry
Laurus nebrascensis Lesquereux
Laurus Newberryana Hollick
Laurus plutonia Heer
Laurus proteaefolia Lesquereux
Laurus teliformis Lesquereux
Leguminosites canavalioides Berry
Leguminosites convolutus Lesquereux?
Leguminosites coronilloides Heer
Leguminosites omphaloboides Lesquereux
Ligustrum subtile Hollick
Liriodendron attenuatum Hollick
Liriodendron morganensis Berry
Liriodendron oblongifolium Newberry?
Liriodendropsis angustifolia Newberry
Liriodendropsis constricta Hollick
Liriodendropsis retusa (Heer) Hollick
Liriodendropsis simplex Newberry
Liriodendropsis spectabilis Hollick
Lycopodium cretaceum Berry
Magnolia amplifolia Heer
Magnolia Boulayana Lesquereux
Magnolia Capellinii Heer
Magnolia Hollicki Berry
Magnolia Isbergiana Heer
Magnolia Lacoeana Lesquereux
Magnolia longipes Hollick
Magnolia obtusata Heer
Magnolia pseudoacuminata Lesquereux
Magnolia speciosa Heer
Magnolia tenuifolia Lesquereux
Magnolia Van Ingeni Hollick
Magnolia woodbridgensis Newberry
Malapoenna falcifolia (Lesquereux) Knowlton
Marsilea Andersoni Hollick
Menispermites acutilobus Lesquereux?
Menispermites Brysoniana Hollick
Microzamia? dubia Berry
Moriconia americana Berry
Myrica Brittoniana Berry
Myrica cliffwoodensis Berry
Myrica longa Heer
Myrica Zenkeri (Ettingshausen) Velenovsky
Myrsine borealis Heer
Myrsine crassa Lesquereux
Myrsine Gaudini (Lesquereux)
Myrtophyllum sapindoides Hollick
Nectandra imperfecta Hollick
Nelumbo Kempii Hollick
Nelumbo primaeva Berry
Ocotea nassauensis Hollick
Onoclea inquirenda Hollick
Osmunda delawarensis Berry
Osmunda novae-caesareae Berry
Paliurus integrifolius Hollick
Paliurus populiferus Berry
Panax cretacea Heer
Periploca cretacea Hollick
Persea Leconteana Lesquereux
Persea valida Hollick
Phaseolites manhassettensis Hollick
Phragmites ? cliffwoodensis Berry
Phyllites cliffwoodensis Berry
Picea cliffwoodensis Berry
Pinus Andraei Coeymans?
Pinus delicatulus Berry
Pinus mattewanensis Berry
Pinus protoscleropitys Holden
Pitoxylon anomalum Holden
Pitoxylon foliosum Holden
Pitoxylon Hollicki Knowlton
Pistia Nordenskioldi (Heer) Berry
Planera betuloides Hollick
Platanus Kümmeli Berry
Podozamites Knowltoni Berry
Podozamites lanceolatus (L. and H.) F. Braun
Podozamites marginatus Heer
Populus stygia Heer
Populites tenuifolius Berry
Premnophyllum trigonum Velenovsky
Protodammara speciosa Hollick and Jeffrey
Protophyllocladus lobatus Berry

Protophyllocladus subintegrifolius (Lesquereux) Berry
Quercus eoprinoides Berry
Quercus Hollickii Berry
Quercus Holmesii Lesquereux
Quercus Morrisoniana Lesquereux
Quercus ? novae-caesareae Hollick
Quercus severnensis Berry
Quercus sp. Berry
Raritania gracilis (Newberry) Hollick and Jeffrey
Rhamnites apiculatus Lesquereux
Rhamnus inaequilateris Lesquereux
Rhamnus novae-caesareae Berry
Rhus cretacea Heer?
Sabalites magothiensis Berry
Sagenopteris variabilis Velenovsky
Salix fluexuosa Newberry
Salix Lesquereuxi Berry
Salix mattewanensis Berry
Salix meekii Newberry
Salix purpuroides Hollick
Sapindus apiculatus Velenovsky?
Sapindus imperfectus Hollick
Sapindus morrisoni Heer
Sapotacites Knowltoni Berry
Sassafras acutilobum Lesquereux
Sassafras angustilobum Hollick
Sassafras progenitor Newberry
Sequoia ambigua Heer
Sequoia concinna Heer
Sequoia fastigiata (Sternberg) Heer?
Sequoia gracilis Heer?
Sequoia heterophylla Velenovsky
Sequoia Reichenbachi (Geinitz) Heer
Smilax raritanensis Berry?
Sphaerites raritanensis Berry
Sterculia cliffwoodensis Berry
Sterculia minima Berry
Sterculia prelabrusca Hollick
Sterculia Snowii Lesquereux?
Sterculia Snowii bilobatum Berry
Sterculia sp. Hollick
Strobilites inquirendus Hollick
Strobilites perplexus Hollick
Thuja cretacea (Heer) Newberry
Thyrsopteris grevilliodes (Heer) Hollick
Tricalycites major Hollick
Tricalycites papyraceus Newberry
Tricarpellites striatus Newberry
Viburnum Hollickii Berry
Viburnum integrifolium Newberry
Viburnum mattewanensis Berry
Widdringtonites fasciculatus Hollick
Widdringtonites Reichii (Ettingshausen) Heer
Widdringtonites subtilis Heer
Williamsonia delawarensis Berry
Williamsonia marylandica Berry
Williamsonia problematica (Newberry) Ward
Zizyphus cliffwoodensis Berry
Zizyphus elegans Hollick
Zizyphus groenlandicus Heer
Zizyphus Lewisiana Hollick?
Zizyphus oblongus Hollick

A CRITICAL STUDY OF THE FOSSIL BIRD *GALLINULOIDES WYOMINGENSIS* EASTMAN

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A number of years ago, in the *Geological Magazine* of London, Dr. Eastman described one of the most interesting and complete skeletons of a fossil bird that we have in this country.[1] This remarkable specimen (see Fig. 1) was taken in the Green River shales of Wyoming, near the town of Fossil, during the summer previous to the publication of the article, and it passed into the possession of the Museum of Comparative Zoölogy of Cambridge (Massachusetts), where it still forms a part of the collection of fossil vertebrata.

During the latter part of May, 1914, I met Dr. Eastman at the United States National Museum, where he was engaged upon a study of the fossil fishes belonging to that institution. He kindly suggested that I communicate with Professor Samuel Henshaw of the Museum of Comparative Zoölogy of Cambridge, and borrow, if possible, the slab containing this fossil bird, and make a more complete study of it than he had made when the specimen first came to him for determination. This suggestion I was glad to act upon; and in a few days, through the courtesy of Professor Henshaw, the original slab came to hand for my study and description.

First of all I made two perfect negatives of the specimen (8′×10′), reproducing it nearly natural size; the reproduction of a print from one of these forms the subject of Fig. 1 in the present article.

In his account of this fossil, Dr. Eastman speaks of it as "a nearly perfect skeleton of a gallinaceous bird" (p. 54), and had he adhered to that opinion, the place he thus assigned it to in the system would never have been questioned. However, he evidently,

[1] Charles R. Eastman, "New Fossil Bird and Fish Remains from the Middle Eocene of Wyoming," *Geol. Mag.*, London, VII (February, 1900), 54–58, Pl. IV (reduced rather more than one-third).

upon subsequent examination, was convinced that he saw other features presented by it, which led him to say, on the same page just quoted, that it was one of an extinct genus of "short-billed, stout-legged birds attaining the size of a gallinule, rail, or small coot, and resembling these forms in general character."

As the supersuborder Ralliformes has no especial affinity with the supersuborder Galliformes, this last statement distinctly contradicts Dr. Eastman's first reference as cited above.[1]

As will be observed by an examination of Fig. 1, the skeleton of this fossil presents many evidences of compression. Most of the bones are of a dark chocolate color, and all of them are considerably darker than the pale-gray matrix in which the specimen is imbedded. Some of the bones have retained their normal positions in articulation, while the remainder have, for the most part, been more or less dislodged from the places they occupied in life. Some were either not in sight at all, or became, to a greater or less extent, removed from the skeleton before the process of fossilization commenced. An excellent example of this is seen in a rib which lies removed from the nearest bone to it in the skeleton by a distance of 18 mm.

This specimen was evidently "cleaned up" by an ignorant collector, and in his misguided attempts to improve it by *scraping*, he has terribly mutilated the most important parts in sight. Thus the skull has been practically ruined, and the cervical vertebrae ground down so as to present merely a longitudinal sectional aspect. Dr. Eastman, in commenting upon this, says: "Depredations of this nature are wholly inexcusable and cannot be too severely censured" (p. 55).

The head and cervical portion of the spine are seen upon direct left lateral view, and are but slightly elevated above the level of

[1] R. W. Shufeldt, "An Arrangement of the Families and the Higher Groups of Birds," *Amer. Nat.*, XXXVIII, Nos. 455–56 (November–December, 1904), 833–57. The relation of these groups is given on p. 852. Dr. Eastman asked me if I did not think there was "some tinamou in the specimen." To which I replied that I did not. As a matter of fact, there is far less tinamou in this extinct fossil bird than there is gallinule. The shoulder-girdle, sternum, and pelvis of a tinamou are entirely different, as may be appreciated by comparing the skeleton of *Nothura maculosa* Temm. with Fig. 1 of the present article. I figured a skeleton of this tinamou in the article here cited (Fig. 3, p. 839).

the matrix. As referred to farther on, the dorsal and lumbar vertebrae are considerably out of position, while all the free ribs are, to a great extent, mixed up together. At least three of the left costal ribs or haemapophyses have retained their places in articulation and are in plain sight; they are the last three of the series on that side.

Passing to the pelvis, we are to observe that it is turned entirely over, so that almost a direct ventral view is obtained of it. Posteriorly, the left femur lies obliquely across it.

Its sacrum is nearly in line with the dorso-lumbar section of the spine; and as to the skeleton of the tail, all but the leading caudal vertebrae are entirely out of sight, although I am inclined to think that the severely ground-down chain of three or four little bones, seen 3.5 cm. to the right of the left tarso-metatarsus, are the partly exposed caudal vertebrae, though the pygostyle is not in view.

The cervical portion of the spine is arched far backward over the dorsum, which is often the case with dead birds found in nature at the present time, and which appears to be due to the mode of attachment of the dorso-vertebral ligaments.

In assuming this position, the skull was likewise carried backward, although its ligamentous attachment to the atlas has been, in part, freed.

Almost a direct left lateral aspect of the sternum is presented, and all this portion is beautifully exposed. Its characters are in plain view, and will be described in full farther on in this article.

The left coracoid has been but slightly dislodged from its coracoidal articular groove on the anterior part of the sternum, and is consequently seen practically *in situ* with respect to the latter bone. This is almost as true of the left scapula, for it occupies nearly a normal position—that is, with respect to its coracoid, the os furculum, and the sternum.

On the other hand, the right coracoid is entirely disassociated from all the remaining bones of the pectoral arch, and now, with its anterior surface exposed, it lies back of the os furculum, in contact with the two clavicular limbs of its arch. We have in view the antero-oblique surface of the furculum, and consequently we have a left latero-oblique side of its hypocleidium exposed.

The right scapula is entirely free, being in full view except the parts covered by the os furculum and the left coracoid which are in front of it, the former almost completely hiding its head. Somehow, its distal head came to get in front of the ends of the vertebral ribs of the left side (see Fig. 1).

With respect to the sternum, it is to be observed that its pair of left xiphoidal processes, as well as the extreme posterior tip of the carina, overlap the shaft of the right femur—the former entirely and the latter as far as the inner condyle.

Perfect in nearly all respects, the right pectoral limb is drawn forward in front of the trunk skeleton, it being crossed by the cervical vertebrae at the distal third of the humerus,[1] which latter has its palmar aspect exposed to view, while it is the anconal sides of the bones of the antibrachium and pinion which are exposed—the forearm and manus having twisted once round before settling down. Curiously enough, the interosseous space between the radius and ulna of this limb is filled in with dark, fossilized structures, as though the muscles of the forearm had become mineralized instead of having been destroyed either by putrefaction or maceration.

Both the ulnare and the radiale of the carpus are in sight, while the skeleton of manus is perfect.

With its longitudinal axis parallel to that of the fellow of the opposite side, the left humerus appears to be free, with its radial side exposed to view.

Passing to the pelvic limbs, they are seen to be nearly perfect, and are so, all to the patellae (which may have been small or not in sight) and some of the pedal phalanges, which will be enumerated farther on. Neither femur was freed at its pelvic articulation, the head, in each case, apparently being still in the cotyloid cavity of the pelvis of the respective sides, as in life, while all the articulations among the long bones below are but slightly out of place in any particular instance.

[1] Dr. Eastman evidently mistook this right pectoral limb for the left. That it is the one of the right side is made almost certain by the fact that the cervical vertebrae would naturally be between the two humeri, when the former was curved backward and the latter drawn forward and upward, as is the case here.

FIG. 1.—*Gallinuloides wyomingensis* Eastman. Reproduction from a photograph of the specimen *in situ*, made direct from the slab by the author; nearly natural size.

On the whole, these limbs are drawn out nearly directly backward, just as though the skeleton had been in running water, the stream passing posteriorward, and these pelvic limbs had, before finally settling down, drifted into the positions in which they were eventually preserved; while the pectoral limbs, influenced by the same current, perhaps periodically flowing to and fro, had previously *lodged* finally in the positions they now are in in their stony matrix. Naturally, the head and neck floated backward with the legs.

The longitudinal axes of the femora are quite parallel to each other; those of the tibio-tarsi are at a slightly open angle, which is also true of those of the tarso-metatarsi, that is, supposing the imaginary lines of their longitudinal axes to be extended toward the pelvis in each instance. The right femur has its anterior surface exposed; the left its latero-internal. In the leg, both fibulae are visible; and in the case of the one on the left side, its proximal head can be plainly seen in the cleft of the external femoral condyle intended for its accommodation and articulation.

Notwithstanding the fact that the distal condyles are very prominent and the intercondylar valley rather deep, the right tibio-tarsus and fibula evidently have their direct posterior aspects exposed, with the fibula next to the median plane. In other words, these two bones, maintaining their articulatory relations, have once rotated over on their longitudinal axis, while the prominence of the distal condyles, over what they naturally possess on this posterior aspect of the tibio-tarsus, has evidently been produced by transverse pressure.

In the case of the left tibio-tarsus, which likewise has maintained its mutual and normal articulatory relations, it presents its antero-externo-lateral aspect.

Either tarso-metatarsus has its almost direct anterior surface exposed, and these are, in each case, thrown but slightly out of place with respect to their articulations with the tibio-tarsi.

The phalanges of the right pes are exposed almost entirely on mesial aspect, and the bones are all present, the joints of the second toe being seen on their dorsal aspects, which, with the exception of hallux, is the case with all the phalanges of the left foot. Here some of the joints are missing, having been broken off and lost. This is

the case with the anterior half of the third phalange of the mid-anterior toe, as well as its entire ungual joint. In the case of the outer toe, all is lost beyond the posterior moiety of the third phalange; finally, the osseous claw of hallux, in the case of this foot, is gone.

All the bones of the hyoidean apparatus are either entirely hidden in the matrix, or else they have drifted away from the skeleton prior to the time of its fossilization. In the left orbit, the circlet of sclerotal platelets are still *in situ* and completely fossilized. The lower mandible is duly articulated, and the bony jaws are tightly closed together.

Dr. Eastman presented a number of accurate measurements of this specimen in his article (p. 56), which were as follows:

TABLE OF MEASUREMENTS

(All *lengths* given in millimeters)

	Mm.		Mm.
Head	48	Manus	46
Scapula	48	Femur	42
Coracoid	27	Tibia	58
Furculum	33	Tarso-metatarsus	34
Christa sterni	58	I. Digit (7, 4)	11
Humerus	47	II. Digit (11, 8, 6)	25
Ulna	49	III. Digit (12, 10, 8, 6)	36
Radius	45	IV. Digit (7.5, 5.5, 4, 4, ? 4)	25

Height of knee-joint (estimated)	90
Total height (estimated)	220

To these I may add some of my own measurements (also in millimeters), which are:

SUPPLEMENTARY TABLE

	Mm.
Length of mandible (approx.)	38
Height of skull (cranium)	20
Longitudinal diameter of left orbit	9
Length of carpo-metacarpus	25
Proximal phalange of index digit	12
Distal phalange of index digit	9
Pollex digit	11.5
Medio-longitudinal length of pelvis	43
Length of pubic bone	32
Greatest width of posterior moiety of pelvis	29
Depth of sternum (anteriorly)	30

The skull.—As already pointed out above, this part of the skeleton has been rendered almost useless through the scraping given it by the collector. Still, there is something to be said of it, for even its outline teaches a little, though in regard to this, Dr. Eastman made no comments whatever, beyond deploring the mutilations which have been committed.

Taken in connection with the mandible, the form of the skull of this fossil bird was distinctly of the gallinaceous type, which will be readily appreciated by comparing it with that of any true and average grouse, partridge, or pheasant. Several of these are shown in previous papers of mine (see footnote 1). In general, the characters are the same in all of these. In the specimen under consideration, the external narial aperture was of moderate size, and possessed the usual elliptical outline; the quadrato-jugal bar was slender and straight, while the postfrontal and squamosal processes were united at their anterior apices as in *Bonasa*. A lacrymal bone was of some size, and had a form such as we see in *Phasianus colchius*,[1] while a nasal beyond it closely resembles that element of the cranium in any average tetraonid that has its maxillary process narrow and delicately formed.

There is a good character in the antero-terminal portion of the superior maxillary or upper mandible, for it has a contour that is strictly gallinaceous in all respects,[2] and is essentially quite different from anything of the kind we find either in the rails or in gallinules.

Vertebral column.—There were fifteen vertebrae in the cervical division of the spine before we come to one which is considered to be the first dorsal vertebra, as it has a pair of true ribs connecting with the sternum by means of costal ribs. This is the sixteenth of the spine, and its ribs and the connecting costal ribs or haemapophyses can easily be made out in the specimen. Whether the fifteenth bore a small pair of free ribs cannot be stated positively, as they are not in sight in that vertebra.

[1] R. W. Shufeldt, "*Osteology of Birds*," State Museum Bull. 130, N.Y. State Museum, Albany, 1909, *Gallinae*, Pl. 2, Fig. 18.

[2] R. W. Shufeldt, "Observations upon the Morphology of *Gallus bankiva* of India," *Jour. Comp. Med. and Surg.*, New York, July, 1888, Vol. IX, No. 4, art. 21, pp. 343–76, 30 figures in text (pmx.).

The dorsal portion of the spine, the thoracic ribs, and the costal ribs are in such a demoralized condition that it would be quite unsafe to make any positive statements in regard to them. There seem to be, however, two free vertebrae just in advance of the pelvis, and each has its centrum rotated into view. In most grouse and other gallinaceous forms there is a single vertebra between the pelvis and the four which coössify into one piece in the dorsal series. If this were the case with respect to the specimen under consideration, and we find, as I say, two free ones anterior to the pelvis, it may be explained by the fact that the fossil skeleton belonged to a subadult individual, which died before the co-ossification of the dorsal vertebrae took place. I do not say that this was the case; but I will say that, were I to find all the rest of this skeleton of this fossil to be typically gallinaceous—and all the dorso-lumbar vertebrae were hidden from sight—its dorsals were four and all in one bone, while between it and the pelvis would be found another single, free vertebra. This would surely be the case were the individual an adult bird.

The ribs were slender, and those in mid-series apparently bore "epipleural appendages," as at least in the case of one rib the process can be seen; they are never very prominent or strong in the *Gallinae*.

In the rails and gallinules the number of cervico-dorsal vertebrae between the skull and the pelvis is greater than in any of the true gallinaceous forms, while in the former the dorsal vertebrae never unite to constitute a single bone.[1]

The pelvis.—Owing to the fact that the ventral aspect of this bone alone is exposed, only its characters upon that aspect can be touched upon. These indicate that the gallinaceous nature of them is very distinctly marked, and that this pelvis would answer for any grouse of average size, such as *Bonasa umbellus* for example. These characters, too, are better shown in such a pelvis as the one possessed by *Centrocercus* than is the bone in such a bird as *Thaumalea picta*. Were its dorsal aspect exposed, in the fossil specimen here being considered, it would closely resemble the pelvis of *Bonasa*

[1] R. W. Shufeldt, "Osteology of *Porzana carolina*," *Jour. Comp. Med. and Surg.*, New York, IX (July, 1888), No. 3, art. 17, pp. 231–48. (Seven figures.) See p. 7.

umbellus, which I figure in my *Osteology of Birds*, published by the New York State Museum at Albany (Pl. VI, Fig. 26).

Anteriorly, the bone is broad, with its iliac borders rounded, the lateral margins being concaved inward toward the sacrum. This latter, on its ventral aspect, presents a longitudinal, median furrow, which extends posteriorly as far back as a point opposite the acetabulae. This, taken in connection with the decided enlargement of the sacrum between the cotyloid cavities and its anterior termination, presents us with a very common character of the pelvis in gallinaceous birds generally. It is well shown in the pelvis of *Centrocercus*. In these gallinaceous birds, too, the anterior sacral vertebra is always prominently produced downward through the form of its centrum, a feature to be noticed in the pelvis of any typical tetraonine species.

Posteriorly, this pelvis is likewise broad, with the free margins of its ilia behind unnotched in any way. The pubic elements are slender in form and of nearly uniform proportions throughout, while posteriorly they extend for some little distance beyond the ilia. In fact, they agree, as do all the other characters of this pelvis, with the corresponding ones in the pelvis of any typical tetraonine species of the present time (Fig. 2).

In the gallinules and other rail-like birds, the pelvis possesses an entirely different character, it being much narrower throughout, with anteriorly truncated ilia and other features, which are clearly to be observed in my figure of the pelvis of *Porzana carolina* referred to above.

Further, I have stated that "in a great many particulars, *Porzana*, *Crex*, *Rallus*, *Ionornis*, *Fulica*, and *Gallinula* agree in their osteology,"[1] and this, in all probability, is true of the remainder of their morphology.

The sternum and shoulder-girdle.—Little need be said by way of description of the sternum of this specimen, for it is so typically

[1] R. W. Shufeldt, "On the Osteology of Certain Cranes, Rails, and Their Allies, with Remarks upon Their Affinities," *Jour. Anat. and Phys.*, London XXIX (October, 1894), N.S., IX, Part I, art. 5, pp. 21–34. Text figures. A paper on the descriptive osteology of all the species constituting this group of birds in North America has recently been accepted for publication by the *Anatomical Record* (Wistar Institute, Philadelphia) and will appear in due course.—R. W. S.

FIG. 2.—Skeleton restoration of the fossil bird *Gallinuloides wyomingensis* Eastman. Made and drawn by the author, guided by the fossil specimen belonging to the Museum of Comparative Zoölogy of Harvard University, kindly loaned him for the purpose. For amount of reduction compare with measurements given in the table above. In some of the bones their actual morphology is closely given. The outline of the skull agrees with the specimen. The cervical vertebrae as drawn are intended to number some fourteen or fifteen. In drawing the pelvis, assistance was given through study of the general tetraonine form of that bone. The free caudal vertebrae and pygostyle are as they occur in related gallinaceous birds. In the case of the ribs, they are in part as in the specimen, while the sternum and bones of the shoulder are entirely so, and agree, in all particulars, with those bones in the specimen of *Gallinuloides* here considered. With respect to the skeleton of the pectoral and pelvic limbs, the bones have the general outline of the corresponding ones, in any particular instance, in the original, and the proportionate lengths are actual.

In making this drawing, I intentionally omitted to include the limbs of the right side, as nothing would be gained by showing them, and they would needlessly complicate the figure. Measurements of the long bones and others are given in the text; and from these measurements, taken in connection with this drawing and other data presented, a very complete conception of the morphology of the skeleton in this fossil bird may be obtained.

gallinaceous in character that similar ones have been described many times in various works on avian osteology. Moreover, with its deep carina; concaved anterior border; a pair of long, somewhat slender, xiphoidal processes on either side, with their expanded, free extremities a prominent manubrial process of quadrilateral form; and finally, a very narrow sternal body, and short costal borders where the costal ribs articulate—all these characters, and a few minor ones, are plainly to be seen in the sternum of this specimen.

This style of sternum agrees in all particulars with that bone as we find it in *Bonasa umbellus*, exhibiting various modifications in the style of sternum found in the genera and families of all gallinaceous birds, in all parts of the world, as for example pheasants, guans, grouse, partridges, guinea-fowls, quails, turkeys, and all their allies and congeners.[1]

Had I seen the sternum of this fossil bird and no other part of its skeleton, I could, without the slightest hesitation, and without having laid eyes upon the coracoids, scapulae, or the os furculum, have described them in all detail. No fowl, living or extinct, possesses a sternum in all respects agreeing with that bone in *Gallinuloides wyomingensis* without having an os furculum which is of the U-shaped pattern, with a large, subtriangular hypocleidium. The clavicular limbs are of nearly uniform caliber, and the free superior ends are but very slightly enlarged. Both *Gallus* and *Bonasa* possess an os furculum of identically the same character, while *Gallinula*, *Fulica*, *Rallus*, and the rest, possess a very different form of one, and one that is more or less compressed antero-posteriorly, without any hypocleidium worth mentioning.

Either coracoid is large, above the average in length for ordinary birds of this size, apart from the gallinaceous group, with the sternal

[1] Richard Owen, *Comp. Anat. and Phys. of Verts.*, II, 27; R. W. Shufeldt, "Osteology of Birds," New York State Museum Bull. 130, p. 183, Figs. 8 and 9 (*Gallus bankiva*); Max Fürbringer, *Ueber Morph. und Systematik der Vögel*, II, Table VI, Figs. 43, 44, 45; on the same plate these may be compared with the sternums of *Fulica*, *Rallina*, and *Crypturus*, in order to exemplify the differences between the galline sternum and the bone as it occurs in rails, gallinules, and tinamous; A. Chauveau, *Comp. Anat. of the Domestic Animals*, p. 113, Fig. 73; T. H. Huxley, *Anat. Vert. Animals*, p. 241, Fig. 81; W. K. Parker, Art. "Bird," *Encyclo. Brit.*, 9th ed., III, 720.

end but moderately expanded, and the head of the bone of a corresponding degree of development.

The inferior or lateral external process is practically very small and this is likewise the case with the "praecoracoidal process" at its superior extremity. With respect to the latter, Dr. Hans Gadow has remarked that this is "but very small or absent in *Apteryx*, *Tinamous*, *Steganopodes*, *Gallinae* and *Passeres*."[1] In its general character, this coracoid agrees closely with that bone as found in the extinct fossil species of *Palaeortyx* and *Palaeoperdix*, as described by Milne-Edwards, Lydekker, Gervais, Depéret, and others.[2]

A scapula is a long, slender, and narrow bone, but slightly curved, and with a small, nib-like expansion at the distal end, which is broken off in the specimen, leaving only its impression in the matrix. In short, the pectoral arch in *Gallinuloides wyomingensis* is quite typically gallinaceous, and essentially agrees with the pectoral arch in the skeletons of existing species and genera of North American *Tetraonidae*, especially with that part of the skeleton in *Bonasa*.

Skeleton of the limbs.—The long bones of the extremities exhibit in a few instances some flattening and slight distortion as the result of pressure. This has not, however, altered or concealed their various characters, but may be clearly made out in most instances. This is particularly the case with the skeleton of the pelvic limbs, which agree, in all respects, with the corresponding bones in the skeleton of an average grouse. When I say this, I refer especially to the tibio-tarsi and metatarsi, which are quite typical. It also had a grouse's wing, in so far as its skeleton goes, as may be easily proved, not only by such characters as have been preserved, but by the relative proportional lengths of the bones of the brachium, antibrachium, and manus. These relative proportional lengths are of some considerable value in making such comparisons, as, within

[1] *A Dictionary of Birds*, by Alfred Newton, assisted by Hans Gadow, with contributions from Richard Lydekker, Charles S. Roy, and Robert W. Shufeldt, p. 857 (footnote).

[2] Compare with the coracoid of *Palaeortyx maxima* of Lydekker in *P.Z.S.* (1893), p. 520, Pl. XII, Fig. 11; see also my figures of the coracoid and the os furculum of *Tympanuchus* in Hayden's 12th *Ann. Geol. Surv. of the Terr.*, p. 716, Pl. XII, Figs. 86 and 87.

certain limits, they hold true for members of the same family, and in several families differ very widely.

CONCLUSION

The extinct fossil bird which has been very fully considered in the present paper was one rather smaller than the North American ruffed grouse (*Bonasa umbellus*), to which it was quite closely affined. Were it in existence today, its place would have been in the family *Tetraonidae*, and near *Bonasa*, *Canachites*, and *Lagopus*, with which forms it holds many, indeed all, tetraonine skeletal characters in common. That it was a bird possessing strong volant powers is abundantly shown by the deep keel to its sternum, and the powerful development of its pectoral arch and wing; these offer ample evidence of this fact. Without doubt it had a flight quite coequal with that of any ordinary grouse.

This form was in no way related to the *Rallidae*, or any similar group, and certainly not to the gallinules. Moreover, when we come to find early Eocene forms of the *Tetraonidae* that exhibit in their skeletons a departure from the true tetraonine stock, it will not be a form of bird having in its skeleton paludicoline characters, and particularly not ralline ones. No such alliances have existed at any time in the world's history.

Annectent forms, possessing a good proportion of tetraonine or galline characters in their generalized organizations, will also exhibit skeletal characters connecting them with the pigeons and their near allies (*Columbae*), and such birds are still found to exist in the world's avifauna, as witness the sand grouse and others (*Syrrhaptes paradoxus*). In fact, the pigeons and fowls, as we know, possess many morphological characters in common at the present day.

As to finding the fossil remains of a true gallinaceous bird in the Green River shales of Wyoming (Middle Eocene), it need not surprise anyone, for true grouse forms occurred in corresponding formations elsewhere, not only in this country but in Europe. This is likewise true of the *Phasianidae*, a matter which I have abundantly proved in a memoir on the subject recently published entitled "Fossil Birds in the Marsh Collection of Yale University," and published in the *Transactions of the Connecticut Academy of Arts and Sciences*, XIX. (February 1915), 1–110.

The generic name *Gallinuloides* is a very misleading one as applied to the present bird, and matters have not been improved by the creation of a family *Gallinuloididae*.[1]

This bird, as I have remarked above, was a true grouse; but whether it can be placed in any existing genus of our North American *Tetraonidae* is another matter. It belongs in the near neighborhood of *Lagopus* and *Bonasa*, and probably in a genus of its own. For such a genus, did the "Canons of Zoölogical Nomenclature" (Canon XXXI) admit of it, I would suggest the name of *Palaeobonasa*. And if at any time in the future such changes are made, under additional rules to meet cases of the kind, this extinct species should then be known as *Palaeobonasa wyomingensis*.[2]

[1] Frederic A. Lucas, No. 4, "Characters and Relations of *Gallinuloides wyomingensis* Eastman, a Fossil Gallinaceous Bird from the Green River Shales of Wyoming," *Bull. Mus. Comp. Zoöl. at Harvard Coll.*, XXXVI (1900–1901), 79–84. Illustrated.

The author of this well-known paper says of the fossil bird here being considered that "its galliform nature is obvious at a glance"; and while he states that it was a "bird of about the size of a ruffed grouse" (which is correct), he falls into the error of stating that "the majority of its structural resemblances are with the curassows and with the genus *Ortalis* amongst those birds," and the still more remarkable error in the statement that "the bird presents no points of affinity with any of the American grouse, still less with any of the *Odontophorinae*."

The writer is very careful to make no reference whatever to the fact that this fossil bird is in no way related to the gallinules; nor does he make any effort (through change of name) to disabuse the mind of the palaeontologists of the incorrectness of that reference. When any animal has been incorrectly classified, it is a distinct advantage to science to rename it, as it is relegated to the group to which it in reality belongs. Moreover, Lucas emphasizes the error made in the original reference by suggesting the creation of the family *Gallinuloididae*, which has, unfortunately, already passed into palaeornithological nomenclature (*A.O.U. Check-List of North American Birds*, ed. 1910, p. 388, etc.).

It may be said here that I figure (natural size) the trunk skeleton of a specimen of *Ortalis macalli* in my *Osteology of Birds*, and give an account of the skeleton of this species (p. 240, Pl. 4, Fig. 21). As compared with the fossil, especial attention is invited to the broad ribs, short and wide external xiphoidal processes of the sternum, and the small hypocleidium of the os furcula in *Ortalis*.

[2] Gen. name = Gr. παλαιός, ancient, and Sp. name, Gr. βόνασος, a wild bull. *Bonasa* was so named on account of the "drumming" practised by the bird at certain times, which has been considered by some to sound like the bellowing of a bull; in other words, an ancient form of *Bonasa*. As I have elsewhere shown, *Bonasa*, of all our North American grouse, comes nearest to the quails and partridges, and it is quite likely that this extinct species possessed some osteological characters in its skeleton which would indicate such an affinity still more plainly.

There have not been many fossil forms of ralline birds described for North America up to the present time, while typical gallinaceous species are, as I have said, by no means rare. Over twenty years have passed since I described *Crecoides osbornii*, a species I then considered as being related to the genus *Crex* among the *Rallidae*.[1] No fossil gallinules have been found anywhere apparently, the nearest form being *Porphyrio mackintoshi*, from the Pleistocene of Queensland, Australia, a specimen which I have not seen.[2] The fossil remains of several extinct forms of *Notornis*, however, have been described, and about five species of *Fulica*, from various and widely separated parts of the world, one of which is from the Pleistocene of Oregon (*F. minor* Shuf.).

[1] R. W. Shufeldt, *Jour. Acad. Phila.*, XI (1892), 412.

[2] De Vis (p. 193, note).

GROUND-ICE WEDGES

THE DOMINANT FORM OF GROUND-ICE ON THE NORTH COAST OF ALASKA

E. DE K. LEFFINGWELL

It is a widely known fact that the ground in Arctic and sub-Arctic regions is permanently frozen to a great depth, only the upper few feet thawing in summer. Nearly all observers have reported the presence of bodies of more or less clear ice underlying the surface of the ground, usually immediately below the limit of annual thawing. Ordinarily the ice is represented as existing in horizontal beds of some thickness and lateral extent, but the observations of the writer upon the north shore of Alaska show that there, at least, the ground-ice occurs chiefly in a network of vertical wedges, surrounding isolated bodies of the tundra formation.

Although this form of ice is the dominant one in the area studied, it is not held that it is the only one, nor that the theory of its formation will fit every case. It seems quite certain that there are several different kinds of ground-ice, each one having originated in a different way.

During the summer of 1914 several dozen photographs of the ice were made, but most of them were damaged later by water, so that the writer has to depend chiefly upon sketches which were often hastily made. Fortunately Mr. P. S. Smith, of the United States Geological Survey, had, some years ago, secured photographs of ground-ice on the Noatak River, and one of these photographs illustrates the wedge-form ice which is the subject of this paper (Figs. 1, 2).

The chief difficulty encountered in the study of the tundra formation arises from faulty exposures. The ground being of material only consolidated by frost, a short exposure to the summer air will cause slumping and consequent masking of the details. It is only where wave or river action has undermined the face of a bank, so that large blocks break off, that good exposures are formed. As

FIG. 1.—Ice wedges, Noatak River. Photograph by P. S. Smith, U.S. Geological Survey.

FIG. 2.—Another view of exposure shown in Fig. 1. Photograph by P. S. Smith, U.S. Geological Survey.

soon as slumping has taken place, erroneous conclusions may be drawn as to the distribution of the ice, scattered outcrops being interpreted as exposed parts of a single bed.

The upper surface of the ground-ice is usually only a foot or two under the surface of the tundra. Consequently in an area which has discontinuous bodies of ice separated by masses of muck, etc., there will be the least amount of material for slumping exactly where the ice occurs. The ice melts back under the overhanging turf, forming a cave, but at either side the muck will slump from

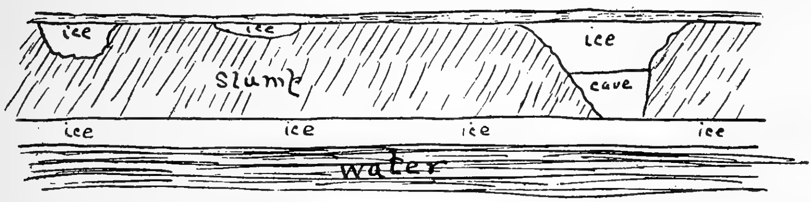

FIG. 3.—Exposure of a bank showing an apparently continuous thick bed of ice

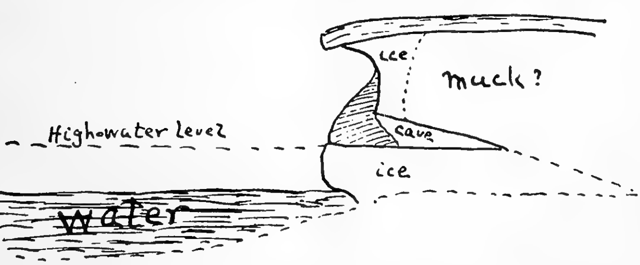

FIG. 4.—Structure of exposure in Fig. 1

the grass roots. Thus, wherever the surface of the bank is exposed, ice is likely to be seen, and observers are led to believe that a continuous body of ice underlies the whole area.

In addition to the erroneous impression as to lateral extent, the conclusions as to thickness are also often faulty. A winter's snow drift against the foot of a bank may be covered by slumping and later exposed, apparently showing a thickness of ice only limited by the height of the bank. The same may be true of river- or sea-ice. An illustration is given of an apparently continuous bed of heavy ice which the writer examined carefully for a quarter of a mile before learning its true nature (Figs. 3 and 4).

The waves had undercut this bank the summer before, making a long, low cave, perhaps ten feet deep. During the following winter this cave had been flooded at high tide and partially filled with ice. Early in the summer the face of the bank had been masked by slumping, leaving only a few glimpses of true ground-ice under the turf. Shortly before the time of observation the lower part of the bank had been washed clear of débris, exposing the continuous layer of new sea-ice. This appeared to be ground-ice, when coupled with the scattered exposures of undoubted ground-ice above. Luckily the cave was exposed to view at one point, so that the mistake in interpretation could be corrected.

The writer went into the field in the summer of 1906 with the idea that the coastal ground-ice occurred in horizontal sheets, and in consequence of faulty exposures did not learn its real distribution until 1914. During the first eight summers, although the ground-ice was examined at every opportunity, little insight was gained into the method of its formation. There has been no opportunity to go anew through the literature before writing this paper, but such illustrations as are at hand do not bear out the inferences drawn from them as to large horizontal beds of ice.

The usual theory advanced in the literature is that bodies of snow or ice were buried by peat or wash material and thus preserved. The writer sought to interpret the Alaskan coastal ground-ice in the light of this theory, but could neither postulate a satisfactory source for the ice, nor find any workable hypothesis to account for its preservation. It was not until the summer of 1914 that the fact was forced upon him that most of the ice was formed in place in the ground. A vertical wedge of ice within a peat bed first drew his attention to the fact; for such a dike of ice could not have stood up in the air for the hundreds of years that were necessary for the formation of the peat (Fig. 5).

FROST CRACKS

During the Arctic winter, frequent reports are heard, coming apparently from the ground. Often the sound is accompanied by a distinct shock, which is in fact an earthquake of sufficient intensity to rattle dishes, etc. One is justified in ascribing this phe-

nomenon to the cracking of the frozen ground during the winter's contraction. The writer has spent six winters in the region under discussion, living most of the time upon the tundra, which is chiefly underlain by muck. Frequent camps have been made upon other formations, such as sands and silts, and the impression carried away is that the sound of the cracking ground was heard everywhere. This has been confirmed by a prospector who has lived nearly thirty years in the country.

It was at first thought that the reports were caused by the cracking of hard snowdrifts, but the fresh cracks in these drifts were seen to run into the ground below. When the snow melts in the summer,

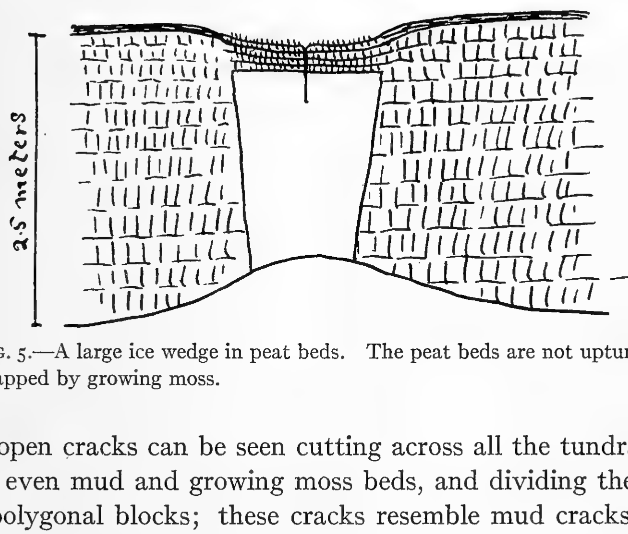

FIG. 5.—A large ice wedge in peat beds. The peat beds are not upturned. The ice is capped by growing moss.

fresh open cracks can be seen cutting across all the tundra formations, even mud and growing moss beds, and dividing the surface into polygonal blocks; these cracks resemble mud cracks but are of a larger size. The blocks have an estimated average diameter of about 15 meters, and have a tendency toward the hexagonal form, although rectangles and pentagons are often developed.

Occasionally a crack is seen to run across a flat surface with no associated features, as is illustrated in Fig. 6, but usually it is accompanied by a distinct topography. Either the crack lies in a gentle depression which surrounds an elevated polygon block, or it runs between two parallel ridges which surround a depressed block. These features do not vary from block to block, but each is locally developed over a considerable area. The "elevated blocks" have

seldom a relief of more than one foot, but that of the "depressed blocks" may be twice as much.

The parallel ridges form shallow reservoirs, very similar to those of the block system of irrigation, especially when they take a rectangular form, as is often the case. They often contain ponds and are always swampy, so that one keeps to the ridges for dry footing when crossing such an area (Figs. 7 and 8).

FIG. 6.—A frost crack on the surface of a recently drained area.

FORMATION OF ICE WEDGES

The open frost crack is in a favorable position for being filled with water during the melting of the snow, for most of them lie in depressions upon a flat surface. Even those that by chance get no water probably become filled with ice crystals deposited by the damp air, by internal "breathing." The crack, being filled with solid ice from the freezing of the water, or containing much ice in the form of frost crystals, thus contains a narrow vein of true ground-ice in the portion which lies below the depth reached by the annual thawing. When the frozen ground expands under the summer's heat, the readjustment to the strain may take place in four ways: (1) The pressure may melt the ice, so that the crack is closed again. (2) The formation

FIG. 7.—A frost crack lying between parallel ridges which inclose depressed polygon areas.

FIG. 8.—Frost cracks, parallel ridges, and block ponds

may be sufficiently elastic to absorb the strain, so that no deformation occurs. (3) The formation may be deformed and bulged up, either as a whole or locally along the edge of the ice wedge. (4) The ice may be deformed.

If the summer's strain has been relieved by readjustment of the material within the polygon block, the next winter will again bring about the conditions which caused the first cracking of the ground. Even if the first crack is full of ice, it may be still a plane of weakness for tensive strains; and this will be especially true if the crack has been only partially filled. Granted that it is a plane of weakness, the new cracks will open at the same place and a constantly growing body of ice be formed at the locus. That this is the ordinary case in tundra formations is seen in the constant association of ice wedges with definite loci of frost cracks.

Thus the growth of the ice goes on from year to year, possibly failing during mild years, when it may not be necessary for all the cracks to open in order to relieve the strain. If the process were not hindered, the upper edges of the wedges would eventually come into contact, thus completely inclosing a cone-shaped mass of the original ground-material. It is conceivable that the process might still go on by bulging up the ice, as it at first bulged up the ground. It is thus within the limits of possibility that a continuous horizontal bed of ice should be formed in this manner, but nothing approaching this possible stage has been observed by the writer.

The thinnest wedge that has come under observation was about a foot wide, but cracks have been seen accompanied by no surface manifestations (Fig. 6) and with no visible ice below them. No doubt the intervening stages exist, especially in an area such as a recently drained lake bottom, where the process is being initiated. The thin veins have nearly parallel sides and flat tops, as can be seen in Figs. 9 and 10. As the ice increases in size, it becomes more wedge-like in form, since the growth is greatest near the top where the crack opens widest. There is a tendency in the large wedges to spread out under the surface of the ground (Fig. 11). This is exaggerated in oblique sections, as is shown in Fig. 12.

The bottom of an ice wedge has never been observed by the writer. Most of the banks on the north shore of Alaska are less

than ten feet high and the bottom is nearly always concealed by slumping. The maximum vertical dimension observed is about 3 meters, but the wedge had a thickness sufficient to have carried it two or three times as far down before pinching out. The ultimate

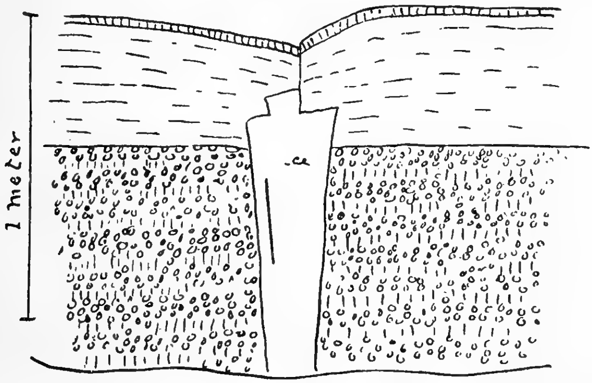

FIG. 9.—A narrow ice wedge in a deposit of mixed clay and ice granules. An open crack within the wedge.

Ground Ice
Lessingwell

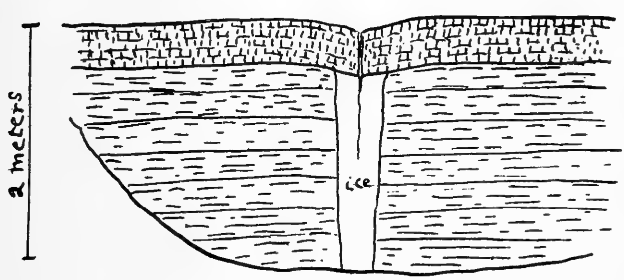

FIG. 10.—A narrow ice wedge in muck beds; an open frost crack runs through the turf and into the ice.

depth must depend upon the depth of annual change in ground temperature in the region. The constants are so uncertain that it is impossible to make a close calculation of this depth. For the purposes of this paper the depth of the ice wedges is assumed to be from 8 to 10 meters.

The upper surface of the ice is usually less than two feet under the ground in muck formations, about the limit to which the summer's thawing penetrates. This surface is usually horizontal, or undulating with the surface of the ground. One or two exposures showed a dome-shaped surface, and another, a central projection above the general surface (Fig. 9). Some more complicated exposures are shown in Figs. 12 and 13. The overlying material is usually muck capped by a few inches of turf. Occasionally it is peat capped by growing sphagnum (?) moss.

Without going deeply into the question of the crystallization of ice, it may be remarked that ice resulting from snow (and glaciers)

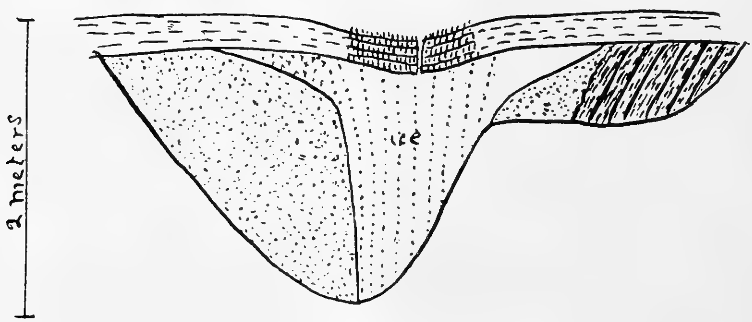

FIG. 11.—A large ice wedge which spreads out under the surface of the ground. The vertical lines indicate rows of whiter ice full of air bubbles. The material on either side is sand. To the right are upturned muck beds.

is granulated, but that from standing fresh water is vertically prismatic. Sea-ice is different from both, but very little has been written upon the subject. The writer's own observations are that sea-ice loses its salt at temperatures approaching 0° C. and becomes honeycombed, showing a general vertical structure, but decidedly different from that of fresh-water ice.

A fresh transverse section of an ice wedge shows a face of whitish ice with numerous parallel vertical markings (Figs. 11–14). These markings are usually of whiter ice which is seen to contain an unusual amount of air bubbles. It is often visibly granular, yet shows a general vertical structure, and breaks up into short, irregular pieces when allowed to melt slowly in the shade.

At the sides of the wedge the markings of the ice are inclined from the vertical and approach parallelism with the sides. Since the growth seems to occur near the center of the wedge, the older lines, though originally vertical, are later spread apart at the top. Oblique sections of wedges will give exaggerated angles of inclination or even curves (Figs. 14 and 15).

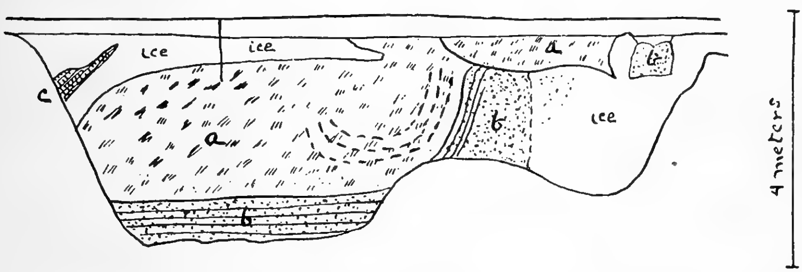

FIG. 12.—A complicated exposure: *a*, disturbed muck and clay; *b*, clay; *c*, peat; at *d* a frost crack runs through the turf and ice.

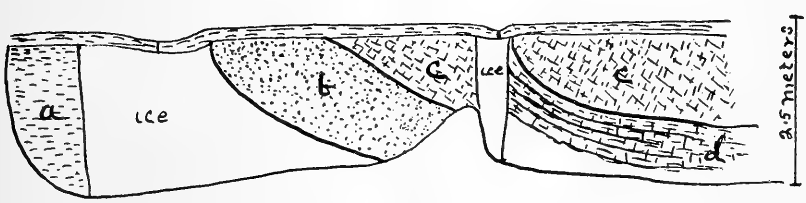

FIG. 13.—A complicated exposure: *a*, clay; *b*, sand; *c*, peaty detritus, no structure visible; *d*, peaty detritus.

In several cases open cracks were seen running down for a few feet into the ice, often being a prolongation of an open frost crack in the tundra above. Once or twice open cracks were seen within the body of the ice, so that a thin sheath knife could be shoved in for several inches (Figs. 9 and 10). Near the edge of a bank these open cracks may become drainage lines for surface waters, so that a tunnel is developed within the ice (Fig. 16). As the tunnel widens the roof caves in and a deep gully is formed in the bank. These gullies work back and around the polygon blocks, making the neighborhood of an old bank rather difficult walking.

The typical formation associated with ice wedges in the region under discussion is muck, a black mud containing much vegetable matter more or less decomposed. It varies from a peaty detritus, which shows signs of having been waterlaid, to sand or mud mixed with a varying amount of decaying vegetation. Undisturbed sections of this muck will usually show horizontal bedding. Occasionally sand or a slimy clay was seen under the muck where a good exposure revealed the lower strata (Fig. 12). As the ice wedge grows in thickness and presses against the edges of the cleaved muck and sand beds, they may become upturned and in time bent to the

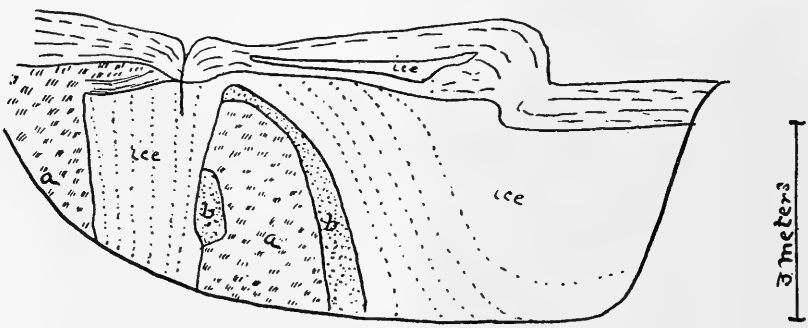

FIG. 14.—Two joining wedges; the one on the right is cut obliquely. The dotted lines represent lines of air bubbles within the ice. *a*, muck and clay, much disturbed; *b*, sand. See Fig. 15.

vertical or even beyond (Figs. 12, 17 and 18), causing the ridges which often run along either side of the frost crack in "depressed block" areas. In "elevated block" areas, the process is not so easily understood. It may be that the block as a whole has bulged sufficiently to bring its surface up to the general level, or else a central depression has been filled by growth and capped by turf.

The writer's observations were insufficient to disclose the factors which control the character of block development. The "elevated blocks" are characteristic of drained areas and are nearly constant features near banks. The "depressed blocks" are associated in the writer's mind with flat, marshy country. This, however, may be the effect rather than the cause of the difference in character of the blocks. The network of depressions drains the elevated blocks, but the ridges form dams which interfere with surface drainage.

As the growing vein of ice becomes more wedge-like in form, the pressure exerts a vertical component against the sides of the wedge. This tends to force the wedge upward. If an upward movement should occur, the ice would carry its protective covering with it and be able to exist level with or even somewhat above the general surface of the block. Since a bulging of the block by the growing wedge seems necessary, some upward motion of the wedge may have taken place without bringing the top of the wedge up to the

FIG. 15.—Photograph of exposure shown in Fig. 14

general level. In the depressed blocks it is to be noticed (Figs. 7 and 8) that the surface of the ground between the parallel ridges (probably underlain by ice) is higher than that of the blocks on either side. No exposures were found illustrating this case, so it is impossible to say whether the surface of the ice is actually higher than that of the blocks.

The usual covering for the ice is muck capped by turf, or peat capped by growing sphagnum (?) moss. As the thickness of this mantle increases by surface growth, the limit of the summer's thawing should rise, thus allowing a constant upward extension of the surface of the ice wedge at the locus of growth. Only two or three cases of apparent upward growth of the surface were seen. In one

case (Fig. 9) there is an upward projection of ice above the general surface of the wedge, indicating a sudden change of the limit of thawing. The others showed a dome-shaped surface, indicating a gradual change. Since the majority of exposures show the surface of the wedges to be nearly parallel with the surface of the ground,

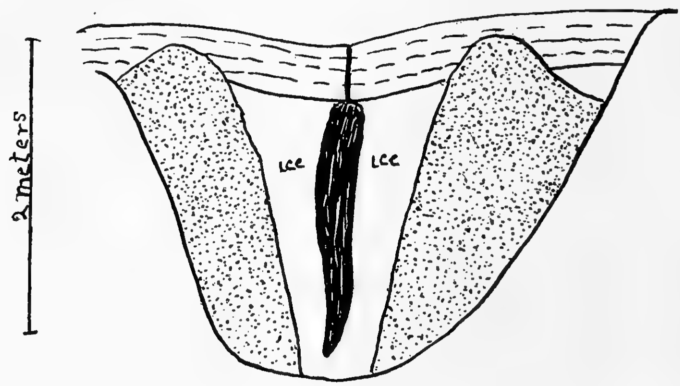

FIG. 16.—An ice wedge in sand. A tunnel has been cut in the ice by drainage of surface waters through the frost crack. The sand on either side is apparently bulged.

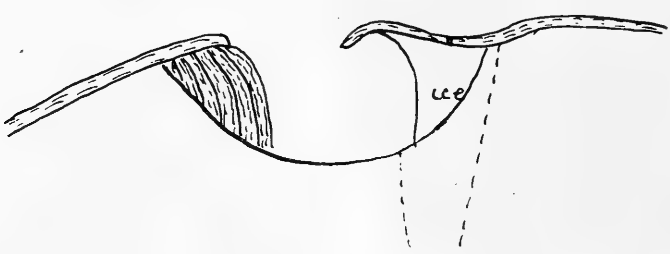

FIG. 17.—Section perpendicular to a bank, showing upturned muck beds in a tundra block which has broken off along an ice wedge.

it seems that a balance must be maintained between the thickness of the covering and the increase in area to be covered, as the wedge becomes wider.

The rate of growth of turf must be very slow in this region, for there are many half-buried bowlders on the surface of the tundra, which have been there since glacial times, or at least since the coastal plain emerged from the sea. Many bare spots exist where the turf has not been able to get a footing.

Where the protective covering is of muck, creeping of the soil will tend to close up the open frost crack. This will thin the covering, and if the rate of surface growth is not sufficient to counteract the resulting decrease in thickness, the upper surface of the ice will be lowered by melting. The increased slopes will cause side material to creep down over the ice, thus keeping the protective mantle up to the required thickness. A shallow depression will thus be formed whose slopes are of the proper angle to cause the proper amount of creeping.

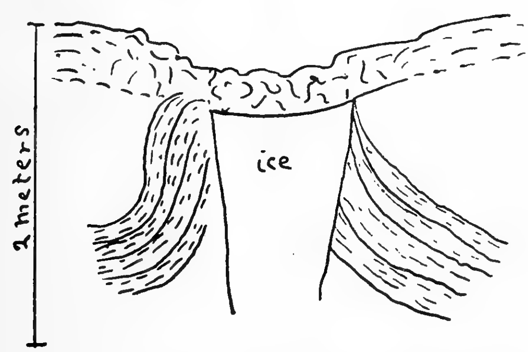

FIG. 18.—An ice wedge in muck showing upturned strata

In the case of sphagnum (?) moss the case may be somewhat similar. The moss and subjacent peat may also close the crack by creeping. At the same time the bed will become thinner, but growth of the moss will soon thicken it again. If the moss grows too fast, the depression will be filled and the conditions of moisture favorable to growth will cease. Thus it is possible for the growing ice wedge to have a covering of peat of constant thickness.

When a bank is undercut by wave or river action, large masses of tundra often break off. Since the ice wedges are planes of weakness, the break is along the edges of the polygon blocks, whenever this is possible. This is especially true of high banks, where whole blocks will break out, leaving many re-entrant angles in the resultant bank, along which a nearly continuous exposure of ice may be seen. The impression is carried away of a heavy horizontal sheet

of ice of great thickness, whereas, in fact, little of it extends back more than a few feet from the face of the exposure (Figs. 19, 20, and 21).

In Fig. 22 is shown a plane-table map of an area of frost cracks, and a sketch of the exposures of ground-ice in the bank immediately below. The polygon blocks were of the elevated type, but the relief was very faint, being somewhat obscured by sand which had drifted up from a sand spit on the left. The exposure was somewhat slumped, but gave sufficient details to illustrate the case.

FIG. 19.—Exposure of ice extending some distance inland. A tundra block is breaking off along an ice wedge.

The heavy lines on the map show open frost cracks; the dotted lines, the evident loci of cracks. Where there was no surface indication, no lines were drawn. The shaded areas are supposedly underlain by ice wedges.

Ground-ice occurs at the intersection of every crack with the bank, where the details were not masked by slumping.

The average diameter of eleven blocks shown on the map is about 11 meters. The largest block is about 11 by 15 meters, and the smallest 5 by 8 meters. The largest wedges of ice are about 2.5 meters wide at the top, and this width has been used in indicating the areas probably underlain by ice, except where surface

indications pointed to new loci. About 20 per cent of the tundra is found to be probably underlain by ice of greater or less thickness.

On a 250-mile boat trip from Flaxman Island to Point Barrow in the summer of 1914, good exposures of muck banks always revealed ice. Many miles were examined closely on foot or from the boat, and very little ice was observed which was not definitely in the form of vertical wedges associated with frost cracks on the surface of the tundra.

FIG. 20.—Tundra block broken off. Ice wedge at left

Ground-ice wedges with their accompanying surface features are typically associated with muck formations, and none were seen elsewhere. River silts, elevated sand and gravel deposits, and soft shales have been carefully examined and the only ice found in such places was evidently of another form and of a different origin. Straight lines leading across the gentle surface undulations of sand spits have frequently been observed, and they could be explained only by frost cracks. No polygonal forms have been seen in such places. The writer is unable to say whether ice wedges develop in such sands, for the exposures made by fresh wave-cutting are seldom more than 2 or 3 feet deep, which is less than the depth reached by the summer's thawing.

RATE OF GROWTH OF WEDGES

Fresh ice-filled cracks 8–10 mm. wide have been observed in the ground immediately above the ice wedges. This may be put as the maximum width of the crack. Open cracks about 5 mm. wide have been found in the ice itself near the upper surface. The width, of course, diminishes downward. If 5 mm. is assumed as the width at the top, it would require only 600 years to build up a wedge 3 meters wide, which is about the maximum width seen in the region.

FIG. 21.—A large tundra mass broken off along the sides of a polygon block. The ice has mostly melted away.

If the cracks do not all open every winter, this period must be multiplied by some factor. The writer had frequently observed open cracks during the previous years, but not realizing their bearing, did not keep any record of their abundance. About 1,000 years seems to be the order of age of the largest wedges. Unless some unknown cause prevents a greater growth, the temperature could not have been sufficiently low to bring them into existence at an earlier date, or else the coastal plain has not been elevated above sea-level for a longer period.

If we assume that the elevated blocks are bulged by the growing ice, the amount of general elevation of the surface of the tundra can

readily be calculated. If 20 per cent is taken for the surface compression of the block, the average compression will be 10 per cent. An average block 11 meters in diameter will be compressed horizontally 1.1 meters to an assumed depth of from 8 to 10 meters. This will cause an increase in a vertical direction of about 1 meter.

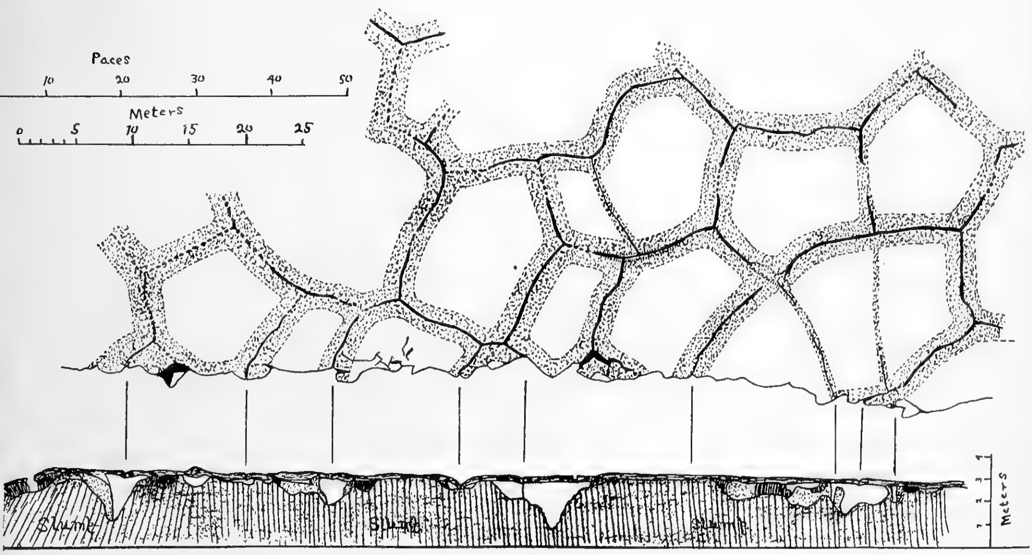

FIG. 22.—Planetable map of frost cracks on the tundra, with a sketch of the exposures of ground-ice in the bank at one edge of the mapped area.

The heavy lines on the map represent open frost cracks in July. The dotted lines indicate evident frost-crack loci. The stipple marks show the areas probably underlain by ground-ice.

In the section below the map, the white spaces represent ground-ice, the dotted spaces, sand. The rest of the exposure has slumped.

The case is different with the depressed blocks, where the adjustment is concentrated into the ridges that surround them. The depressions are being continually filled with growing vegetation, as well as picking up wind-blown material, thus forming muck beds. In this way a much greater general elevation of the surface of the tundra is possible. Much of the muck of the region may have been developed in this way.

The principle of the development of ground-ice wedges is capable of widespread action throughout the region of permanently frozen ground. It is so persistent on the north shore of Alaska that it is to be expected to come into play in similar regions elsewhere. The writer is inclined to believe that much of the ground-ice described in the literature as in horizontal sheets may turn out to be in vertical wedges. In the classical locality at Eschscholtz Bay, no one observer agrees with the others. One says that there is a solid mountain of ice, while a second finds only a thin veneer of ice against the face of the bank. A more careful observer finds scattered outcrops of ice, including at least one vertical dike. In other regions "polygon marks" and vertical markings upon whitish ice are mentioned.

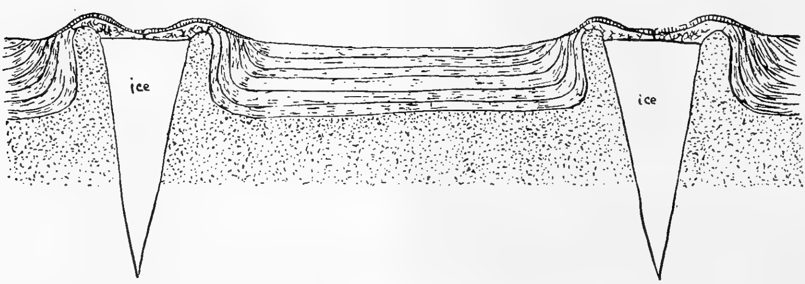

FIG. 23.—Hypothetical section of ice wedges and depressed polygon block

SUMMARY

The permanently frozen ground contracts in the cold Arctic winter and cracks are formed, which divide the surface of the ground into polygonal blocks. In the spring these frost cracks become filled with surface water which immediately freezes. In the expansion of the frozen ground as its temperature rises in summer, the vein of ice being more rigid than the country formation, the readjustment takes place in the latter. The result is to bulge up the inclosed block either bodily or else locally along the sides of the ice. During the next winter's cold wave, a new crack forms at the same locus, so that a continually growing wedge of ground-ice is formed (Fig. 23). Thus the tundra becomes underlain by a network of ice wedges, which inclose bodies of the original formation.

STRATIGRAPHY OF THE WAVERLY FORMATIONS OF CENTRAL AND SOUTHERN OHIO[1]

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PART I

PART II

[1]Published by permission of the State Geologist of Ohio.

PART I

CLASSIFICATION

The following columns of formations (Table I) in as many different regions are presented chiefly to indicate the members composing the Cuyahoga in its different facies. To these have been added the other formations with some duplication of names so that the columns are somewhat more than a classification; they are a partial correlation table. The top of the Cuyahoga on the Ohio River, as explained later, is the stratigraphic equivalent of the lowest part of the Byer member farther east and north. The amount of equivalency cannot be exactly indicated.

The names Vinton, Allensville, Byer, Berne, Fairfield, Lithopolis, Churn Creek, Vanceburg, Rarden, Henley, and Portsmouth are new formational names,[1] and the name Buena Vista is applied somewhat differently from its original usage but not with a wholly new meaning. The name Raccoon is adopted from Hicks.[2]

In presenting a summary of the findings of several years' work in this field it is impossible to cite any but the chief facts. The presentation of the large body of evidence on which these conclusions rest, soundly it is believed, must await the completion of the work now being prosecuted on the faunas of the region. Such

[1] The names Vinton, Allensville, and Byer were first used without definition in a chapter by the writer entitled, "The Geological History of Fairfield County, Ohio," in the *History of Fairfield County*, pp. 203–23 (Chicago: Richmond-Arnold Publishing Co., April 15, 1912). The names were applied to the respective members in Fairfield County but they were not defined. They are here defined for the first time.

[2] L. E. Hicks, *Am. Jour. Sci.*, 3d. Ser., XVI (1878), 216.

TABLE I

Central Fairfield and Hocking Counties		Central Licking County		Eastern Licking County	
Logan formation	Vinton member Allensville member Byer member	Logan formation	Vinton member Allensville member Byer member	Logan formation	Vinton member Allensville member Byer member
Cuyahoga formation, Hocking Valley conglomerate facies	Berne member Black Hand member Fairfield member Lithopolis member	Cuyahoga formation, Granville shale facies	Berne member Black Hand member Raccoon shales, top only seen, rest known from wells	Cuyahoga formation, Toboso conglomerate facies	Berne member Black Hand member "Sandstones" (wells) "Shales" (wells)
Sunbury shale Berea Grit Bedford formation		Sunbury shale (wells) Berea Grit (wells) Bedford formation (wells)		Sunbury shale (wells) Berea Grit (wells) Bedford formation (wells)	

Southwestern Scioto County		Central Ross, Pike, and Scioto Counties	
Logan formation not subdivided		Logan formation not subdivided	
Cuyahoga formation, Vanceburg sandstone facies	Churn Creek member Vanceburg member Rarden member Buena Vista member Henley member	Cuyahoga formation, Scioto Valley shale facies	Portsmouth member Buena Vista member Henley member
Sunbury shale Berea Grit Bedford formation		Sunbury shale Berea Grit Bedford formation	

relationships as would necessitate even brief discussion of facts for and against alternate interpretations have been passed over.

It is obviously impossible even to mention the results obtained by many previous workers who have contributed, often fundamentally, to the subject. The stratigraphy of the upper part of the Waverly had been studied in detail in a single small region only, in central and eastern Licking County, near the center of the state, and at the northern end of the area herein considered. Hicks in 1878 gave a brief description of the formations of this limited region and Herrick in the eighties described other subdivisions, particularly with reference to the faunas. Prosser[1] in 1901 reviewed the work of previous writers on the stratigraphy, adopted the subdivisions made by Hicks and Herrick, applied to them the proper geographical names, and described in detail some of the sections in the region. Prosser's description of the sections at Newark, subdivided after Hicks and Herrick, is the one generally quoted in more or less detail as representing these formations in Ohio.

It is not surprising to find that, as a result of the changed perspective afforded by detailed work over a much wider area, the subdivisions adopted by Prosser in this section must be revised. In fact, both Hicks and Herrick failed to correlate correctly the members across this small area in Licking County which they studied in some detail. In addition, it appears that one member, described as a formation, the Black Hand, is not as important stratigraphically as some members that were not particularly distinguished by the earlier workers.

In Table II are given the subdivisions of the Newark section as stated by Prosser, and in a parallel column the ones herein adopted for the same section. This will serve to show the extent to which the interpretation of the section has been altered.

It will be noted that the names Cuyahoga and Logan are retained for two chief terranes, and these are extensively subdivided into members. These have been recognized as two chief terranes of

[1] *Jour. Geol.*, IX (1901), 205–31. The Quarry Run section is described on pp. 221–26. Other outcrops seen later led to a second description of this section with certain changes in interpretation. This was published in 1904, *American Geologist*, XXXIV, 358–61.

the Waverly for forty years. When the errors of interpretation and correlation of various geologists during this length of time are resolved out, it is peculiar that, after all the different applications that have been made of these names, it should now fall to them to denote precisely the same beds that they were originally applied to forty years ago, or, as nearly as can be determined, the central Ohio equivalents of these beds. The subdivisions of the Logan here adopted constitute exactly the Logan formation at Logan, the locality from which E. B. Andrews named it in 1870, since when the

TABLE II

Proposed Classification		Prosser	
Logan formation	Vinton member		Logan formation
	Allensville member	Conglomerate II	Black Hand formation
		Allerisma shale	
	Byer member		
Cuyahoga formation	Berne member	Conglomerate I	
	Black Hand member		
	Raccoon shales, only a few feet exposed		Cuyahoga formation

name Logan has been used with precisely the same meaning by no other writer. The Cuyahoga was named from outcrops on the Cuyahoga River in northern Ohio where it is overlain by the Pennsylvanian, the whole of the Logan and some of the Cuyahoga having been removed by pre-Pennsylvanian erosion. It is not, at present, possible to demonstrate that the beds to which the name Cuyahoga is here applied are the precise equivalent of the beds in the type section; but it is now evident that they are almost or quite identical with those shales in the shale facies in central Ohio which Orton knew lay between the Berea Grit and the Logan, which he called Cuyahoga, but of which he never fixed the precise limits.

This does not imply that any of these workers understood the true relationships. They did not; none of them knew that the Black Hand (or Waverly conglomerate as it was called) and the underlying coarse sandstones (which indeed were practically

unknown) are merely a facies of the Cuyahoga, and the attempts, particularly by Orton, properly to place these conglomerates in the Waverly column led to confusion. He included the Black Hand with Andrew's Logan under the term Logan Group, a procedure that has been followed by many.

Yet, in spite of the fact that these names are now made to denote the same beds that they denoted in the earliest usage in central Ohio, a question may be raised by some whether they should be retained; whether it is established that these subdivisions are two chief subdivisions, and that a classification which would drop one or both terms and place some of the members here proposed in a higher rank would not more nearly indicate the true relations. Prosser in his work in northeastern Ohio has dropped the term Cuyahoga[1] and builds this portion of his Waverly column out of members which were formerly units within the Cuyahoga. From this it would appear that he does not regard the Cuyahoga shale, as delimited by Newberry in northern Ohio, as a terrane of established stratigraphic significance.

Apropos of this question it may be remarked that in the present work it has been found exceedingly convenient to retain both. The classification of the Cuyahoga of central and southern Ohio here proposed with its numerous members in different facies is very complex; nevertheless, when the entire area is considered, it appears that these groups of members constitute a distinct terrane, for no one of its members in any facies can be satisfactorily discussed except in connection with the other members of that facies. Furthermore, the Berne member, whether it be regarded as closing the Cuyahoga or opening the Logan, separates two groups of sediments that are essentially different from each other in many ways. The retention of the term Logan is necessary, for although easily subdivided into three important members over most of this area, along its southwestern margin it is not so divisible and one name is needed to denote sediments that are there the equivalent

[1] "The Devonian and Mississippian Formations of Northeastern Ohio"; *Geol. Surv. Ohio,* 4th ser., *Bull. 15* (1912), 574 pp.

of all three a few miles to the eastward. It is true, as will be pointed out, that on the Ohio River the lower part of the Logan sandstones pass into shales that are, so far, indistinguishable from the upper part of the Cuyahoga and are included in the upper part of the Cuyahoga. Nevertheless, for the reasons just stated, the conceptions of Logan and Cuyahoga are exceedingly convenient ones, whatever may be their actual stratigraphic importance to be determined later. Final decision should be reserved until the evidence of the faunas can be presented.

Perhaps the greatest innovation from the Waverly column as heretofore established is the reduction in rank of the Black Hand formation. This had come to be regarded as a distinct formation between the Cuyahoga and Logan, a stratigraphic unit that had to be considered in any attempt at correlation of rocks of this general age. It is, however, a local development of the Cuyahoga and is, apparently, not as important a stratigraphic unit as any one of the three members of the Logan.

BEDFORD AND BEREA FORMATIONS

In the classification of the Waverly formations here presented, the Bedford formation has been included. By such inclusion it is not intended to express any opinion in the discussion now being waged as to whether this formation should be placed in the Devonian or should be retained in the Waverly as was the practice for many years; it is here retained largely because such has been the practice. The present discussion does not attempt to correlate this or other subdivisions of the Waverly with the standard time scale because that would involve discussion of the faunas, a subject not yet ready for presentation. An earlier paper by Hyde on a special phase of the Bedford and Berea gives their general character in southern Ohio although by no means a summary of the various problems connected with them.[1]

One feature there mentioned should be emphasized. Since that paper was published several papers have appeared describing the unconformity between the Bedford and the Berea in northern and

[1] *Jour. Geol.*, XIX (1911), 257–69.

FIG. 1.—A. Map showing outcrop belt of Waverly formations of central and southern Ohio, bounded on the western side by Devonian and Silurian rocks, on the eastern side by the Pennsylvanian. These boundaries are drawn only approximately and are generalized. The conglomerate and sandstone facies of the Cuyahoga are stippled, at the northward the Toboso conglomerate area, near the central portion the Hocking Valley conglomerate facies, and in the southwestern part the Vanceburg facies; intervening are the Granville and Scioto Valley shale facies. Question marks indicate uncertainty due to removal by erosion, burial beneath glacial deposits, or unobserved areas. The known western margin of the Allensville member of the Logan is represented by a line east of the center of the Waverly belt. Near the Ohio River its actual stratigraphic western limit is represented by a continuous line. From there northward to Hocking County its known westernmost occurrences are represented by coarse dashes—the horizon occurs possibly as far west as the dotted line and its disappearance westward is known in northern Scioto County, but the precise zone of disappearance is not located. From Hocking County northward it extends as far west as the westernmost outcrops of its horizon, represented by short dashes.

central Ohio.[1] In view of the debated age of the Bedford, such an unconformity, if widespread, would be urged, and indeed it has already been suggested by Prosser,[2] Girty[3] and Burroughs[4] as evidence for the separation of the Bedford from the remainder of the Waverly and of its affiliation with the Devonian. It is therefore important to emphasize that there is no evidence seen by the writer south of Lithopolis in Fairfield County of the existence of any such plane of unconformity. There is irregularity at the plane which in some sections appears to be this contact, but apparently of the same kind and of no more importance than many other similarly irregular bedding planes, often in the same section, and not of a nature, so far as observed by the writer, to warrant description as evidence of erosion. On the other hand, at many localities, there is a gradation from the Bedford to the Berea, sometimes abrupt, elsewhere gradual. The Berea of southern Ohio is only a phase of the Bedford; they consist of precisely the same kind of sediments with the same ripple-marked structure, a sedimentary structure so peculiar and unusual that its occurrence in both Bedford and Berea cannot be urged as an insignificant coincidence by anyone at all familiar with sandstone formations and their structures. Almost the only distinguishing feature between them is that the sandstone beds are

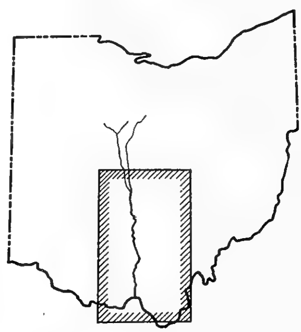

Fig. 1.—B. This outline map is a guide map to indicate the position in Ohio of the area mapped in the large one.

[1] W. G. Burroughs, "The Unconformity between the Bedford and Berea Formations of Northern Ohio," *Jour. Geol.*, XIX (1911), 655–59; C. S. Prosser, "The Disconformity between the Bedford and Berea Formations in Central Ohio," *ibid.*, XX (1912), 585–604; C. S. Prosser, "The Devonian and Mississippian Formations of Northeastern Ohio," *Geol. Surv. Ohio*, 4th ser., *Bull. 15* (1912); W. G. Burroughs, *Economic Geol.*, VIII (1913), 480–81; "Berea Sandstone in Eroded Cleveland Shale", *Jour. Geol.*, XXII (1914), 766-71.

[2] C. S. Prosser, *Geol. Surv. Ohio*, 4th ser., *Bull. 15* (1912), 511, 512.

[3] George H. Girty, "Geological Age of the Bedford Shale of Ohio," *Annals New York Acad. Sci.*, XXII (1912), 296.

[4] *Jour. Geol.*, XXII, (1914), 771.

relatively thicker and the shales relatively thinner in the Berea, and some, not all, of the sandstones of the Bedford are slightly calcareous, never enough so to be called an impure limestone. Furthermore, the same structure in the Berea sandstone generally in central Ohio and at the only one of the localities seen by the writer, Lithopolis, where the unconformity is present, shows that the Berea above the unconformity in central Ohio is of much the same type as, and is at least closely related to, the Berea of southern Ohio, and presumably is not widely different from it in age.

SUNBURY SHALE

The Sunbury shale is a thin bed of black, fissile shale present over the entire area here considered, either on outcrop or under cover. Of it no more need be said in view of Prosser's excellent description of its features,[1] except that it with the subjacent Berea has been invaluable as a base from which to pursue work in the complex overlying formations.

CUYAHOGA FORMATION

THE FACIES

The Cuyahoga formation has for years been described as a shale formation. It is true that sandstone members, sometimes of great economic importance, have been known to occur in it at various points, and thin sandstones in greater or less abundance have always been recognized as a component throughout its thickness. However, the descriptions which have appeared in print have generally been such as to give the impression that it is largely a body of shales and not a great deal more complicated than the formations below it. This is owing to the fact that portions of the massive sandstone and conglomerate facies were not recognized as Cuyahoga but were considered a distinct horizon, the Black Hand formation, or, under the name Logan conglomerate, were considered a part of the overlying Logan formation, while the remaining portions of these same sandstones were practically unknown.

The Cuyahoga formation may be made up very largely of shales in one area but of sandstones or conglomerates in adjacent areas

[1] *Jour. Geol.*, X (1902), 262–312.

at no great distance. Furthermore, in any such area, the sediments are likely to remain much the same from the bottom of the formation to its top. That is, in a sandstone or conglomerate area we may expect to find sandstones or conglomerates or sandy shales more or less continuously throughout the formation, and in a shale area the sandstones are likely to be thin or inconspicuous. This is due to localized conditions of sedimentation which prevailed during Cuyahoga time. The result is that distinct depositional facies can be recognized with pronounced uniformity in the lithology of the vertical column in any one area, except in the regions of transition from one facies to the adjacent one.

It is necessary to recognize five such facies. Beginning in Licking County and passing southward across the outcrop (see outline map), these are (1) the *Toboso facies*, an area in which the sediments are largely non-fossiliferous conglomerates and sandstones with shales in the lower part; (2) the *Granville facies*, mostly shales with finer-grained fossiliferous sandstones in the upper part, conglomerates absent or limited to a few thin beds; (3) *Hocking Valley conglomerate facies*, mostly conglomerates and coarse sandstones with some shale, fossils wanting; (4) *Scioto Valley shale facies*, argillaceous shales with thin sandstone beds tongued into them from the provinces on either side, fossiliferous; (5) the *Vanceburg sandstone facies*, an area in which are developed sparingly fossiliferous sandstones of a different type from those in the conglomerate areas.

Within these facies, the Cuyahoga is composed of distinct lithological members, which can be traced in regular succession over much or all of the facies and sometimes into the adjacent facies. It is these relationships that have given rise to the complex classification here presented.

There is also to be noticed, before taking up the facies in detail, the fact that they all have a pronounced axial trend from north-northwest to south-southeast, and that this axial trend does not correspond to the direction either of the strike or of the dip of the series, but is intermediate between them. As a result the area of any one of these facies extends diagonally across the outcrop belt of the Cuyahoga.

The northern limit of the present work in Licking County is by no means a logical one; it cuts through the heart of the Toboso conglomerate facies. Besides, conglomerate areas have been reported from Knox County to the northward which may be at about the same horizon and in which certain structures are reported which are also found in the Toboso and Hocking Valley provinces.[1] This conglomerate may be at a higher horizon, as its base appears to be about 965 feet above the base of the Bedford shale. Whether they are a continuation of the conglomerates of the Toboso province or whether they belong to a distinct facies is not certainly known, but apparently they are distinct. Other conglomerates of a yet more uncertain horizon are reported from yet farther north in the northeastern part of Richland County,[2] southern Ashland,[3] and in Wayne County[4] where G. F. Lamb describes two conglomerate horizons.

The two conglomerate areas considered in this paper, the Toboso and Hocking Valley, have much in common. Lithologically and in the structure of their beds they are very closely related, and both have clearly been formed in much the same way. The other sandstone facies, the Vanceburg, is wholly different in these and in other respects. Furthermore, the Granville shale facies lying between the two conglomerate areas, although lying mostly below drainage and hence inaccessible except by well-records, is clearly influenced largely by their proximity. The Black Hand and Berne members lying at the top of the Cuyahoga in the conglomerate areas possibly in places under cover extend entirely across the Granville facies, although the former in part passes into shale in its surface outcrops. In fact, it seems very probable that the Toboso and Hocking Valley conglomerate areas are merely lobes extending northwestward from a much larger conglomerate area lying under cover of the Coal Measures at no great distance to the southeastward, and that the Granville shale province is the accumulation of shales with considerable sandstone lying between these lobes. The Scioto shale facies is not influenced to nearly such an extent by the adjacent sandstone areas.

[1] *Geol. Surv. Ohio*, III, 337, 338. [2] *Ibid.*, pp. 317–18. [3] *Ibid.*, p. 523.

[4] *Ibid.*, p. 539; also G. F. Lamb, *Ohio Naturalist*, XIV (1914), 344–46.

THE TOBOSO CONGLOMERATE FACIES

Extent and thickness.—The conglomerates of this facies are exposed typically in the east-central part of Licking County. The areal extent is unknown to the writer, since it has not been traced to the northward, and observation has been confined to the excellent outcrops seen within two or three square miles in the vicinity of the post-glacial gorge of the Licking near Toboso and Hanover. It seems very likely that it is small and not at all comparable with the great Hocking Valley area. The entire observable width is less than three miles, since the Black Hand passes below drainage immediately east of Toboso, as a result of regional dip to the eastward.

The thickness of the Cuyahoga in this area is about 588 feet (gas well at Everett glass-sand plant above Toboso where the upper 340 feet are reported as "sandstone," the remainder as "shale"). Of this, only the topmost 100 feet are exposed above drainage. It was to this exposed 100 feet of conglomerate in the gorge of the Licking that Hicks gave the name Black Hand, from Black Hand Rock near Toboso. From the original Black Hand of Hicks, the topmost foot or two are now separated by the writer as the Berne member.

The Black Hand member.—This member of the Toboso facies is a massive, coarse quartz sandstone with abundant quartz pebbles which are seldom over an inch in diameter and usually half an inch or less. It is pure enough to be used for glass sand after washing. The color is usually yellow or buff, sometimes reddish. Fossils, worm trails excepted, have not been recorded.

The structure of this member is of significance. The sandstones within 10 or 15 feet of its top lie horizontally or nearly so. Below this the bedding of the remainder of the exposed portion, 55 or 60 feet, is inclined, in general, to the northward at angles of from 5° to 15°, 10° being usual. On the western side of the area, that is, near and to the southward of Hanover, the dips are slightly northwest, the strikes ranging from due east and west to N. 80° E. Farther east, and observed especially on the Licking River, the dips swing to east of north, the strikes ranging from N. 80° W. to N. 50° W. This difference in the dips on the east and west sides of

the area within short distances suggests strongly that this conglomerate mass is not one of great extent, and that the more easterly outcrops are really on the eastern side of the conglomerate area. That these are true beds is shown by the occasional occurrence of ripples on their upper surfaces or of a very thin parting of clay shale between two inclined sandstone beds. Such a feature may be found along a bedding-plane from the point of its emergence near stream level to its disappearance near the top of the inclined beds.

Although the inclination of the larger beds of the region is quite regular, the structure within these beds presents the greatest diversity. The material is very coarse sand with abundant pebbles, and evidence of strong current action is seen. Cross-bedding is everywhere present in the inclined portions. These minor structures follow no appreciable order. The cross-bedding is in any direction. It may be in opposite directions in superjacent beds and often is inclined directly against the slope of the larger beds, contradictory as it may seem. Erosion planes are abundant; although local and limited in extent, these are true local erosion planes which can frequently be traced for many yards along an outcrop, and appear to be distinct from the small cuttings which are always found where there is evidence of current action and which are usually referred to simply as cross-bedding. To the eastward the Black Hand goes under cover with these structures yet present, but on the western side of the conglomerate area the bedding gradually flattens out and the current structures and cross-bedding disappear together with the pebbles, so that within a mile or two the member becomes a more or less structureless (except for ordinary bedding-planes), coarse sandstone. This is the way it occurs at Clay Lick only two and one-half miles west of where the structures are at their maximum of development. And only a short distance west of Clay Lick, thin clay shales begin to appear below the massive upper 100 feet or Black Hand member. These suggest that the Granville shale province is being approached and also fairly well demonstrated that the upper 100 feet more or less of sandstones in the Toboso province constitute a member distinct from the sandstones below it reported in the well.

Hicks[1] noticed the general northerly dip of these beds and the oblique lamination dipping in all directions and considered it the structure of a sea beach. No worker since has paid any attention to this acute observation or its significance, and Herrick derived the material from the northeast, flatly contrary to the evidence of the structure.[2]

The Berne member.—At the top of the conglomerates is found, at all points, one or two feet of coarser, more pebbly conglomerate. This is designated the Berne member. Its discussion is reserved to the consideration of the Hocking Valley conglomerate area.

Origin of the conglomerates.—It is evident that the sandstones and conglomerates of this area were derived from the south-southeastward. It is true that only the upper sixth of the thickness is exposed, but it appears, from the structures observed in the Hocking Valley conglomerate area, that the statement may safely be applied to the entire thickness. It appears that the visible portion of the conglomerates at least were built forward by currents of considerable strength, either as a sand-spit, as was suggested by Hicks, or delta-wise. The dips are low, far below the angle of repose for subaquatic accumulation of such materials, and this, together with the exceedingly irregular cross-bedding within the inclined beds, suggests that the material must have been in part distributed by submarine currents.

THE HOCKING VALLEY CONGLOMERATE FACIES

Extent and thickness.—The conglomerate area traversed by the Hocking Valley has been the more closely studied. It lies far enough west on the outcrop belt that the most of its structure can be observed, whereas the Toboso area lies so near the eastern margin of the Waverly belt that only the upper 100 feet of the Cuyahoga are shown.

This area occupies the central and western parts of Fairfield and the western half of Hocking counties, barely extending into Vinton County. The southwestern part of Licking and the eastern margin of Franklin should also be assigned to it. Along its western

[1] *Am. Jour. Sci.*, 3d ser., XVI (1878), 217.

[2] *Bull Sci. Lab. Dennison Univ.*, II, Part 1 (1887), 9, 10.

margin it has a length of 45 or 50 miles, its maximum width is about 20 miles. In the southeastern and eastern portions only the upper beds are exposed; in the western and northwestern portions only the lower part remains. The entire thickness is not exhibited in any one section. However, well-records have been easily obtained from the Sugar Grove gas field which extends across it, and have been of much service in unraveling the structure, although little more than generalized impressions can be obtained from the reports coming from the heavy drills there used.

The sandstones of this province are developed along a very distinct axis which trends north-northwest, south-southeast. Along this line of maximum sandstone development, the thickness is 625 feet, decidedly greater than in the Granville shale province to the eastward, and more than twice as thick as it is some miles to the southwestward in the Scioto Valley shale facies where it is about 295 feet. This increased thickness is due to the increased amount of coarse material.

Four members are recognized in the Cuyahoga of this area.

The Lithopolis member is best exposed in the northwestern part of Fairfield County in the vicinity of Lithopolis. The name is newly proposed. It is a series of *thin, horizontal, interbedded sandstones and shales.* The fact that they are horizontal is to be emphasized. The sandstones are usually light gray or bluish in color, moderately fine grained, and evenly bedded. The beds are usually several inches and sometimes two or three feet thick. The shales are argillaceous, and usually, but not always, somewhat sandy; they are commonly gray in color. Carbonaceous material may be present at times sufficient to darken the gray. The sandstones and shales, on the whole, are about equal in amount, although at various horizons one may dominate the other and the sandstones appear to be considerably in excess in the upper part.

The thickness varies from 118 feet to at least 140 feet and possibly 180 or 200 feet between Lithopolis and Chestnut Ridge two miles to the eastward. It is overlain at these localities by a massive coarse, yellow sandstone which is wholly different from those of the Lithopolis member. To the southward, beneath the central portion of the Hocking Valley province, the Lithopolis

member is entirely below drainage but the well-drillers invariably report from 100 to 200 feet or even more of "shale" at the base of the Cuyahoga. This appears to be largely the Lithopolis member, but the records of drillers are usually difficult to interpret, and it does not follow that the entire "shale" bed reported is Lithopolis. The fact that it is reported as shale also need not be disconcerting, as a series more than half made up of sandstone might be so reported. Nor is the reported variation in thickness necessarily a fact; this interval is never measured by the drillers and is almost always given from memory.

Prosser and Cumings have described the Lithopolis occurrences in minute detail and also similar strata from the lower part of the Cuyahoga in eastern Franklin and western Licking[1] counties. Those of the two last localities have not been examined by the writer, but they appear to be of exactly the same type as those at Lithopolis. They have separated off the lower 50 feet of the beds here called Lithopolis and correlated them with the Buena Vista member, or lower 50 feet of the Cuyahoga according to Orton. Since there is no member in central or southern Ohio so limited which is of any significance, stratigraphic or otherwise, the name Buena Vista has been limited in the present work to the principal quarry stone at Buena Vista in the Vanceburg facies from which it was originally derived, and its stratigraphic equivalent cannot as yet be detected and indeed is probably not recognizable in the Hocking Valley conglomerate area. The stratigraphic horizon of the Buena Vista member, as hereinafter defined, will fall in the Hocking Valley facies at least as high as the middle of the Cuyahoga and possibly higher.

The Fairfield member is a series of alternating sandstone and shale beds which quite frequently show initial dips to the northward of two to six degrees. The sandstones when typically developed are coarse, reddish yellow, brown, or bluish gray, sometimes pebbly, and quite commonly are found in massive members 20–60 feet thick with intervening shaly members of similar thickness. The shale strata, however, are themselves formed of thin interbedded sandstones and shales, the former likely to be very coarse,

[1] *Am. Geol.*, XXXIV (1904), 335–58.

even when thin. The features which characterize the member are the *alternating coarse sandstones and shale, the former often massive and thick*, and the *low initial dips*. At some localities there are no shales and the bedding is not always inclined. The last seems particularly true in the northern part of the area. The member ranges from 200 to 330 feet in thickness, but exact figures are impossible.

The Black Hand member is the principal cliff-forming member of the Cuyahoga, and the beautiful scenery of Fairfield and Hocking counties is chiefly the result of its resistance to erosion. It is a massive coarse sandstone or conglomerate and is distinguished from the underlying Fairfield member by the *absence of shale* (very rarely single beds an inch or two thick may be observed) and the increased prominence of its structural features, notably the *steeply inclined bedding* (*10° to 20°*), the numerous erosion planes, abundant cross-bedding, etc. Its thickness is usually about 100 feet, but may reach 150 feet over the center of the province.

Areally, this member extends laterally beyond the lower members, whenever it can be followed far enough to determine the point. It persists to the eastward with complete loss of structure except normal horizontal bedding into the eastern part of Fairfield County until it passes below drainage, where it is frequently reported in wells with "shales" below it. It is found capping the hills almost as far west as Amanda (Fairfield County) and Tarleton (Pickaway County) but beyond that it is removed by erosion. It is not found more than three or four miles north of Lancaster, the hills beyond that point falling away to lower elevations.

In the southwest corner of Hocking County (Salt Creek Township) and the northeast corner of Ross, and in the western townships of Vinton County, the horizon of the Black Hand is so low in the hills that its passage westward, first into sandstones and then into the shales of the Cuyahoga, can be traced with ease. A marine fauna appears in the sandstones just east of where they pass into shale. The transition has been traced at several points, sufficient to show that the zone of transition extends a distance of 15 miles in a south, slightly southeasterly direction from near the extreme northeast corner of Ross County to near Raysville on the

south line of Vinton County. It has not been traced into Jackson County where a few miles farther to the southward it must pass below drainage.

Over the center of the province, the Black Hand is a coarse pebbly sandstone with occasional beds of conglomerate. The pebbles, all quartz, seldom exceed an inch and a half in diameter. The most striking feature of the member is the bedding, which is inclined at angles ranging commonly from 10° to 20° and sometimes reaching 25° or even more. This structure is much more prominent on the eastern side of the area. This inclination is found throughout most of the thickness of the Black Hand, only the topmost 10–25 feet lying horizontally. The inclined beds correspond in a way to the foreset beds of a delta, the horizontal beds above to the top-set beds. The inclination is almost always toward the northward or at least with a northerly component.

When all the features of the Black Hand and Fairfield members over the area of the facies are taken into consideration, it is evident that the members are distinct. Yet it may be difficult to draw the line between them in any given section, especially in the southern part of the area in western Hocking County. It is impossible here to discuss the relation of one bed to the other. Owing to the physical nature of the members, the contact is almost always covered except for short exposures of a few feet at wide intervals. Where the members are well developed there is almost invariably an erosion plane present at or near this contact. But whether this is a plane of unconformity between the two members or is merely another of the commonly observed erosion surfaces, several of which can sometimes be seen in a single outcrop and most of which are obviously local, is not yet fully determined. In any event the erosion plane, if there is an extensive one, disappears near the margin of the facies as the coarse sediments give place to shales, and is believed to be of no great significance. Whether these conglomerate facies be regarded as deltas of a peculiar type or as bar formations, the surface of the accumulation at any given time stood well above the mud floor of the adjacent shale facies and was liable to erosion over portions of its surface under conditions which would not affect the adjoining deeper areas.

G. F. Lamb has described two conglomerate horizons in Wayne County in northern Ohio.[1] Both have the bedding inclined to the northward, and erosion surfaces are recorded beneath each and are described as unconformities. The lower bed is from 2 to 45 feet thick and is about 625 feet above the Berea sandstone. From this it would appear to be near the horizon of the Black Hand member. It may, however, be the Berne member (which in places is Herrick's Conglomerate I recorded by Herrick in northern Ohio) which was also preceded by erosion over the central Ohio conglomerate areas, and which extends areally beyond the Black Hand both in the Granville facies and on the western side of the Hocking Valley facies.

The Berne member, never over 20 feet thick, and frequently only one foot in thickness, is always present and readily recognized, resting on the Black Hand. The term is proposed from numerous outcrops in Berne Township of Fairfield County. In the Hocking Valley region, as in the Toboso region, this bed consists largely of pebbles, but sandstones of moderate coarseness and shales are found in it at some localities. In mapping, its upper limit forms the logical plane to be followed since it usually forms the top of the ledges and the bed is lithologically very like the underlying conglomerates. The contour of the surface of this bed, that is, the surface of the Cuyahoga deposits, is much the same as that of the Black Hand member, but in order to understand the Berne member it is desirable that the top of the Black Hand be distinguished and that the contour of its surface be determined separately from that of the Berne member, thin though the latter be.

The topographic map is essential to the determination of these features, and at the time the field work on this region was done the Lancaster and Logan sheets were the only ones issued covering any considerable portion of the area. The Chillicothe sheet which lies southwest of the Lancaster sheet, covers only a very small area of the transition zone on the extreme western side of the province. Much of the most interesting part of the province in southern Hocking and the western part of Vinton counties lies south of the

[1] "Middle Mississippian Unconformities and Conglomerates in Northern Ohio, *Ohio Naturalist,* XIV (1914), 344–46.

Lancaster sheet and has not been visited since 1910, when it was unmapped.

On the eastern side of the main body of the facies, that is, that portion of it lying west of the Hocking River, the top of the Black Hand dips to the eastward at a rate which varies considerably from point to point, but is much greater than the general regional dip. In north-central Hocking County the dip over much of this side is about 35 feet per mile, but reaches a rate of 119 feet for a distance of a mile on the eastern flank, just west of Hocking River. A few miles north of this, the dip is about 37 feet per mile over the center of the conglomerate area, increasing to an average of 58 feet over most of the eastern flank, but with no indication of the extreme dip observed for the short distance farther south. Along the eastern margin of this steep slope the surface of the Cuyahoga is carried down nearly to the level of the main drainage lines, but after continuing for a short distance thus, it recovers and rises to the eastward in a broad low arch a few miles wide, within which cliffs of Black Hand 40 or 50 feet high are exposed; beyond this to the eastward it sinks rapidly below drainage. This broad low arch is exposed for four or five miles along the Hocking River above Logan and again along Rush Creek in Hocking County several miles to the northward, as well as along the intermediate smaller streams. It is very like a broad, low anticlinal fold with its axis trending in a generally north-south direction. However, it does not appear probable that the doming of the surface is due to gentle regional warping, although this possibility has not been disproved. It seems more likely that this dome is a second smaller conglomerate mass lying a short distance east of, and nearly contiguous to, the main mass, and is, indeed, merely a lobe of the main Hocking Valley area. The structures observed in the exposed portions bear this out, the material being conglomerate with northerly dips. It probably dies out north of Rush Creek, for no trace of it was observed farther north and east of Lancaster where the sandstone loses its pebbles and structures immediately east of the Hocking River, and the surface declines regularly to the eastward. The contour of the upper surface of the Black Hand is thought to be due to the influence of this separate lobe of accumulation, for

the thickness of the entire Cuyahoga is greatest over this lobe and the main conglomerate area and diminishes somewhat in the slight sag between them.

The principal considerations in support of this are: (1) the lack of evidence of any pronounced anticlinal structure in the underlying rocks, although the region has been drilled over for years in the search for gas and oil; (2) the increased thickness of the Berne member in the trough and its diminished thickness over the crests; the same is true to some extent of the next succeeding bed, the Byer member of the Logan.

On the western side of the Hocking Valley province, the inclination of the top of the Black Hand (and with it the Berne member, which is uniformly thin) becomes considerably lessened and the member may even sink slightly to the westward, notwithstanding the general regional rise of the older rocks in that direction. At the Rock House, about 11 miles west of Logan and just south of the south edge of the Lancaster sheet, the elevation of the top of the Cuyahoga, that is the top of the Berne member, is 1,060 feet above sea level. Eight and one-half miles almost due southwest of this point on the Ross-Hocking county line, the same horizon occurs at only 975 feet, a decline of 85 feet contrary to the general regional dip. The decline in the surface of the formation is here clearly due to the westwardly decrease in thickness of the Cuyahoga.

The Berne member differs very decidedly from the conglomerates lower down in the presence of a marine fauna. This is seldom abundant, either in species or in individuals, but the fossils are so widespread, both in the finer beds and in coarse conglomerates, that they may be considered an essential character.

In the region of the Hocking Valley facies, it is found in its simplest form over the central and western portions. Here, with a few exceptions, it is a more or less massive coarse sandstone with abundant pebbles, on the whole always decidedly coarser than the Black Hand, and usually from one to three feet thick. It becomes thinner westward so that it is usually about one foot thick on the western margin of the region. It even extends westward for a few miles beyond where the Black Hand becomes shaly and in

eastern Ross County is found resting on typical Cuyahoga shales, itself reduced to a horizon of a few inches of soft clay shale or clay ironstone with many small quartz pebbles imbedded therein. It does not pass into sandstone before its final disappearance; it is a pebble bed up to the point of its disappearance.

The Berne member attains its greatest known thickness in the trough between the two domes of conglomerates mentioned above, where, near Enterprise in Hocking County, it is at least 16 and probably 20 feet thick. It there consists of many thin beds of fine pebbles alternating with thin beds of sandstone of varying coarseness, the whole suggesting aggradation of different kinds of material brought by continually shifting currents from the more exposed, shallower portions of the adjacent conglomerate masses to the more sheltered deeper water between them. The member, here and over its entire area, can best be interpreted as the concentrate of pebbles and coarse sand which resulted from the reworking of the top of the earlier deposits by wave and current action. This must have happened soon after the completion of the deposition of the underlying conglomerates and while they were but slightly consolidated.

It is not certain whether the Berne member should be considered the topmost member of the Cuyahoga or the base of the Logan formation. Lithologically it falls readily with the Cuyahoga conglomerates, and in the event of mapping the region it will prove a most convenient upper boundary. There is very good reason to believe, however, that from the point of view of historical succession it belongs with the Logan. This conclusion is supported by several considerations. (1) The base of the bed is sharp, and in places is undoubtedly a plane of erosion as where, near Logan, the horizontal beds of the top of the Black Hand have been removed and the Berne member truncates the inclined beds of the Black Hand. (2) There is occasionally a transition from the Berne member to the Logan although never more than a few inches thick, as in the railroad cut at Hanover (Licking County) and at localities in Hocking County. (3) Beds very like the lower member of the Logan, although usually somewhat coarser, are frequently present in the Berne member east of the Hocking Valley province. (4) The

bed is a marine conglomerate, as shown by the general occurrence of marine fossils in it. Although the body of the Cuyahoga conglomerates is not to be considered as typically non-marine, it is evident, from the lack of all fossils except trails, that it was a very specialized type of marine deposit. The Berne member marks the resumption of conditions under which marine life could exist. (5) The faunas found in the Berne show much closer relationships to those of the Byer member of the Logan than to those of the Black Hand in its marginal transition phases. These faunas do not, however, appear to indicate an essential difference in age. The evidence goes to show that the Black Hand and Logan faunas were contemporary and their distinctness is due to facial differences. The fauna of the Berne member thus indicates the existence of faunal conditions like those of the superjacent Logan rather than those of the subjacent Black Hand.

In concluding the consideration of the Hocking Valley region, it remains to be said that the source of its material must have been in the same direction as that of the Toboso province. The gentle northeastward inclination of the sandstone beds of the Fairfield member on the east side, their occasional westerly dips on the west side, the bedding of the Black Hand sharply inclined to the east, northeast, north, or west, when considered with the southeasterly trend of the axis and the shale areas to the east and west, clearly indicate that the source of the material was to the southeast.

The material of the Cuyahoga sandstones in both the Hocking Valley and the Toboso provinces is quite pure quartz sand and pebbles. Close examination of the crushed rock from the Black Hand shows some small amounts of feldspar and more abundant kaolin, the latter clearly indicating the once greater percentage of the former, but it appears that the material has been much worked to bring it to its present degree of purity. It is true, however, that the samples so examined were selected for their purity in connection with work on glass sands. In the Fairfield member and less rarely in the Black Hand, beds may occasionally be found in which the kaolin is so abundant as to be readily detected in the field without the aid of a lens. These beds must have been quite high in feldspar originally.

THE GRANVILLE SHALE FACIES

Extent and thickness.—This facies lies between the Toboso and Hocking Valley conglomerate facies and is an area in which shales accumulated to a much greater extent than in the other two. It is of unusual interest because it was in the Licking County outcrops of this facies that Herrick worked out certain of his faunal horizons and from which he described many species; furthermore the succession of formations found there, as interpreted by the workers of the time, has been adopted during the past decade as the typical Waverly column of Ohio.

The boundaries of the facies are only partially determined and must be largely inferred at present. In general it appears to occupy most of the eastern half of Fairfield and the western part of Perry counties, where it is almost entirely below drainage, and the central part of southern Licking County. The Toboso conglomerate area lies to the eastward in Licking County and, presumably, also in Perry County, but deep under cover in the latter. The Hocking Valley conglomerate area lies to the westward in central Fairfield County, and its hypothetical northward extension in northern Fairfield and western Licking counties, since removed, is supposed to have limited the Granville province in that direction. To the southeastward, it is partially limited by the small conglomerate lobe on the eastern side of the Hocking Valley area, and it probably does not extend much farther in that direction under the Coal Measures cover before it is bounded by coarse sediments extending continuously from the Toboso to the Hocking Valley facies. It has not been traced to the northward and northwestward, but it probably extends across the northern part of Licking County and somewhere in that direction it formerly united with the shale deposits of the Cuyahoga which, from theoretical considerations, must have extended around the northern end of the Hocking Valley conglomerate facies and must have been continuous with the shales of the Scioto Valley shale facies.

Unfortunately the three northern townships of Fairfield County are almost without exception very heavily drift covered, a buried pre-glacial valley; the northwestern portion of Perry County, although underlain by rock, shows very few outcrops, and at the

most, only the topmost beds of the Cuyahoga are shown. In addition, Licking County southwest of Granville is practically unexamined by the writer.

Wherever this facies has been observed, either in the outcrops or by means of well-records, its sediments show the effect of the conglomerate areas on either side. Their thickness, while somewhat less than in the adjacent conglomerate facies, is much greater than in the Scioto Valley shale facies, and the sandstone content appears to be decidedly greater. As has been stated, the thickness of the Cuyahoga in the Hocking Valley facies is from 600 to 625 feet. Passing eastward this gradually decreases in the wells to about 500 feet just west of Bremen in Fairfield County (Myers well). A few miles to the northward at Rushville, an approximate determination with several somewhat uncertain factors gives 535 feet. At Newark, yet farther north in Licking County, well-records indicate about 570 or 575 feet, and at the Everett quarry over the Toboso facies, due east of Newark, it is 588 feet thick. This series of observations extends in an irregular oblique line north and northeasterly, entirely across the Granville province. Newark, from various considerations, appears to lie on the transition to the Toboso facies.

The Black Hand member appears to be generally present along the length of this line, although there are often unknown covered intervals of several miles for which no well-records or outcrops are known. It appears to be from 50 to 100 feet thick and is considerably changed from its conglomerate phase; it consists largely of coarse or moderately coarse sandstones, with occasional shaly beds, and sometimes carries fossils. Just east of Rushville, there is good evidence from wells that the Black Hand is largely represented by what the drillers report as shale.

Due westward from Newark a change in the nature of the Black Hand can be observed in the surface outcrops. Within four miles it passes largely into sandy shales; sandstone beds are present, sometimes two or three feet thick, but they are irregular and may pinch out entirely in a few feet. This passage to shales exactly parallels that observed at many points on the western side of the Hocking Valley facies in Hocking, Vinton, and Ross counties.

Conditions of outcrop, and limited observation, however, have not permitted the tracing of this passage farther in Licking County. The nature of the Black Hand horizon at Granville, two or three miles yet farther west, is not known, but a further disappearance of the sandstones is anticipated. In fact, so firm is this conviction that no hesitation has been felt in adopting the name Granville for the shale province although very little of the Cuyahoga was seen by the writer in that vicinity. Four or five miles yet farther to the westward on Moot's Run, the fossiliferous beds of the Cuyahoga, there largely a shale but with considerable sandstone, furnish an excellent collecting ground; whether they are Black Hand, shale facies, or belong to the underlying Raccoon member is not yet determined.

Raccoon member.—Below the Black Hand near Newark there are some 20 or 30 feet of sandy or clay shales with numerous thin sandstones, sometimes quite fossiliferous; the remainder of the Cuyahoga, here and other points known to the writer, is below drainage and is commonly reported as shale by the drillers, but an occasional more careful record shows that the next underlying 200 feet carry many beds of sandstone, sometimes 3 to 6 feet thick, with thicker shale beds. The lowermost 200 feet, more or less, appear to consist almost wholly of shale. In 1878, when Hicks proposed the name Black Hand, he used the name Raccoon for the shales between it and the Sunbury shale. When, later, the central Ohio and northern Ohio Waverly formations were correlated, it appeared that this term was synonymous with Cuyahoga and it was dropped. It may be brought forward, temporarily at least, to designate that almost unknown thickness of shales and sandstones below the Black Hand member. So little is known of these at present in the Granville facies that it cannot yet be safely urged as a permanent formational name.

The Berne member.—Little need be said of this member in this province. It is a pebble bed of essentially the same composition as over the two conglomerate provinces, except that beds of finer-grained sandstone are frequently intercalated. Between the two conglomerate facies there is an interval of 20 miles in northern Fairfield and southern Licking counties across which its presence

has not been shown because of drift-covered outcrops or its occurrence below drainage, but it is probably present. Followed westward from the Toboso province it can be traced readily into the Conglomerate I of Herrick in the section in Quarry Run at Newark. At the Dugway three miles west of Newark, where the Black Hand member is largely shaly, the Berne member is 10 feet thick and the pebbles are noticeably smaller than to the eastward.

The section in the Quarry Run at Newark has become the classic section of the upper Waverly of Ohio. Prosser's description of this section[1] in which he adopted the stratigraphic units used by Hicks[2] and Herrick has been accepted as the standard column of the upper Waverly, and most of the correlations that have been made with the upper portion of the Waverly have been based thereon. It is, indeed, the only section of this portion of the scale which has been accurately described. It appears, however, that the Black Hand has been made to include too much in the descriptions of this section. Only the lower 60 feet are the equivalent of the Black Hand as that member is interpreted in the present work. Conglomerate I of Herrick, which was placed by Prosser 29 feet below the top of the Black Hand, is the Berne member and the top of the Cuyahoga. These 29 feet are to be correlated with the Byer member and the lower part of the Allensville member of the Logan. Prosser's error is the result of his having accepted Hicks's statement that all of these beds were the equivalent of his (Hicks's) Black Hand at Clay Lick and Toboso. To Hicks belongs the credit of having first appreciated the passage of the Black Hand conglomerates into finer-grained sediments to the westward. If Orton had appreciated the significance of Hicks's discovery, much of the confusion of later years would have been spared.

[1] *Jour. Geol.*, IX (1901), 221–26; *Am. Geol.*, XXXIV (1904) 358–61.

[2] *Am. Jour. Sci.*, 3d ser., XVI (1878), 216.

[*To be continued*]

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NOVEMBER-DECEMBER 1915

A REVISION OF THE SEQUENCE AND STRUCTURE OF THE PRE-KEWEENAWAN FORMATIONS OF THE EASTERN GOGEBIC IRON RANGE OF MICHICAN

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AND

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Assistant Geologist, Michigan Geological Survey

INTRODUCTORY STATEMENT

This paper is a brief presentation of the main results of recent studies of the geology of the Gogebic iron range in Michigan from T. 47 N., R. 45 W., east to Gogebic Lake, a distance of 16 miles.

Former interpretations of the geology are based almost entirely on early investigations of the United States Geological Survey[1] recently summarized in *Monograph 52*. Prior to this work four years of study of the pre-Cambrian of the region south and east of Gogebic Lake to the Iron River district, including 38 townships adjacent in northern Wisconsin, by the senior writer and his assistants had failed to establish a satisfactory basis of correlation with the rocks of the Gogebic Range. Inasmuch as recent drilling for iron ore has supplied important information, unavailable to

[1] C. R. Van Hise and R. D. Irving, *Monograph 19, U.S. Geol. Survey.*

earlier workers in this field, it was thought that a field study of the eastern Gogebic iron range would probably alter former interpretations of the geology and furnish a basis of correlation with the pre-Cambrian terranes of adjacent territory.

The results of the seasons' studies are more satisfactory than we had anticipated and are of considerable importance to the progress of pre-Cambrian geology of the Lake Superior region. This is our excuse for presenting them in preliminary form in advance of a more thorough treatment.[1]

We acknowledge our indebtedness to Dr. C. R. Van Hise for placing in our hands his early field notes and plats, to Dr. C. K. Leith for helpful suggestions in field conference, to Mr. Robert Selden Rose, of Marquette, and to the Presque Isle Mining Company for records of drilling and underground exploration.

SUMMARY OF THE GEOLOGY BASED ON EARLIER WORK

Van Hise and Leith have recently presented the views of the United States Geological Survey[2] which may be briefly summarized.

The Huronian rests on the Archean with profound unconformity and is represented by two unconformable sedimentary groups, the Upper and the Lower Huronian, which are in approximate structural parallelism and dip steeply northward beneath the Keweenawan.

The Archean comprises a green schist series (Keewatin) intruded by granite (Laurentian).

The Lower Huronian has two members, a basal quartzite (Sunday) and an upper cherty dolomite (Bad River).

The geology of the "Upper Huronian (Animikie) Group of the Eastern Area" is summarized in the following words:

> In the eastern part of the district—that is, from about 6 miles east of Sunday Lake to Gogebic Lake—the Upper Huronian rocks have an exceptional character. In the larger part of the district the conditions were those of quiet sedimentation, but in the eastern area throughout the greater part of the Upper Huronian there was continuous volcanic action. In consequence the

[1] R. C. Allen and L. P. Barrett, "Contributions to the Pre-Cambrian Geology of Northern Michigan and Wisconsin," *Publication 18, Michigan Geol. Survey* (in press).

[2] *Monograph 52, U.S. Geol. Survey.*

rocks are lava flows, volcanic tuffs, conglomerates, agglomerates, and slates, with all sorts of gradations, just such as one would expect if a volcano rose in a sea and volcanic action continued for a great period. Naturally in this area it is not possible to map any continuous sedimentary belts. The dominant rocks are greenstone conglomerates and lavas and massive eruptives. The uppermost formation for the extreme eastern part of the area is ferruginous slate. This ferruginous slate, though dominantly clastic, contains narrow bands of non-clastic sediments such as chert, cherty ferro-dolomite, ferro-dolomitic chert. It is believed that the ferruginous slate is probably at the same horizons as the Ironwood formation to the west and that its dominant fragmental character is due to the presence in this area of one or more volcanic mountains which arose above the water and upon which the waves were at work after the close of the period of active volcanic outbreaks.

SUMMARY OF THE WRITERS' CONCLUSIONS

The writers' most important conclusions, in so far as they differ from those of the earlier writers as summarized above, may be stated briefly as follows:

1. *Huronian.—There are three unconformable sedimentary groups in the Huronian series.* The formations heretofore included in the Upper Huronian (Animikie) group are divisible into two groups separated by an unconformity of the first magnitude. For the superior group we propose the name Copps, in recognition of the important exposures of its basal horizons near the old Copps mine.

2. *Archean.—The granite heretofore described as Laurentian not only intrudes the Keewatin but also cuts across two great unconformities into the Animikie group.* Some of the granite may be Laurentian, but we have not been able to so correlate any of it.

3. *The unconformity between the Copps and Animikie groups is one of great angular and erosional magnitude* and is of greater importance in this district, measured by criteria for evaluation of breaks in pre-Cambrian sedimentary successions, than either the unconformity separating the Keweenawan and Huronian or the one separating the Animikie and Lower Huronian.

For convenience of reference and greater clearness in the discussions which follow we introduce the successions of the United States Geological Survey and the writers' in parallel columns.

CORRELATION TABLE I

Gogebic Iron Range

	By U.S. Geological Survey, 1911		By Michigan Geological Survey, 1914
	West of center of T. 47 N., R. 44 W.	East of center of T. 47 N., R. 44 W.	East of T. 47 N., R. 45 W.
Quaternary System Pleistocene series	Glacial deposits	Glacial deposits	Glacial deposits
Cambrian System		Unconformity Lake Superior sandstone	Unconformity Lake Superior sandstone (doubtful)
Algonkian System Keweenawan series	Unconformity Gabbro, diabase, basic lavas, conglomerate, etc.	Unconformity Gabbro, diabase, basic lavas, and interbedded sandstone and conglomerate	Unconformity Gabbro, diabase, basic lavas, interbedded sandstone and conglomerate and basal conglomerate
			Unconformity Copps formation. Ferruginous slate, graywacke, chert, and basal conglomerate
Huronian Group Upper Huronian (Animikie)	Unconformity Greenstone intrusives and extrusives Tyler slate Ironwood formation (iron-bearing) Palms formation	Unconformity Extrusive and intrusive greenstones Palms and Ironwood formations represented by ferruginous slate, graywacke, and quartzite	Unconformity Granite Extrusive and intrusive greenstones Ironwood formation (iron-bearing) Palms formation
Middle Huronian Lower Huronian	Unconformity Bad River limestone Sunday quartzite	Absent	Unconformity Bad River limestone Sunday quartzite
Archean System Laurentian series	Unconformity Granite and granitoid gneiss	Unconformity Granite and granitoid gneiss	Unconformity
Keewatin series	Greenstone and greenschist	Greenstone and greenschist	Greenschist

ARCHEAN SYSTEM

Rocks of Archean age are confined, in the territory under discussion, to the fine-grained chloritic schists (Keewatin) lying unconformably beneath the Sunday quartzite in T. 47 N., R. 44 W. In the maps accompanying the reports of the United States Geological Survey the hornblende and mica schists lying south of the Huronian sediments in T. 47 N., R. 43 W., are included in the Keewatin, and the extensive area of acid intrusives, here termed the Presque Isle granite, is placed in the Laurentian. We believe that the Presque Isle granite intrudes the Middle Huronian and that the schists in T. 47 N., R. 43 W., heretofore correlated with the Keewatin, are phases of the Palms formation, the metamorphism being induced by granitic intrusion.

THE ALGONKIAN SYSTEM

HURONIAN SERIES

Lower Huronian group

The Lower Huronian comprises two formations, viz., the Sunday quartzite and the Bad River limestone, exposures of which are limited to the south half of sections 17 and 18, T. 47 N., R. 44 W.

The Sunday quartzite is grayish white to pink, about 150 feet thick, and carries a thin basal conglomerate holding fragments of the underlying Archean (Keewatin) greenschist. It inclines gently (30°) across the almost vertical plane of schistosity in the subjacent schists.

The Bad River formation is a cherty dolomite about 400 feet thick. It is conformable through gradational phases with the underlying Sunday quartzite and is unconformably overlain by the Upper Huronian (Animikie) group.

Upper Huronian (Animikie) group

The Animikie group is represented by a basal quartzite (Palms), an iron-bearing member (Ironwood), volcanic conglomerates, breccias and flows, and intrusive granite and greenstone.

On recent maps of the United States Geological Survey the Palms and Ironwood formations are carried eastward as cartographic

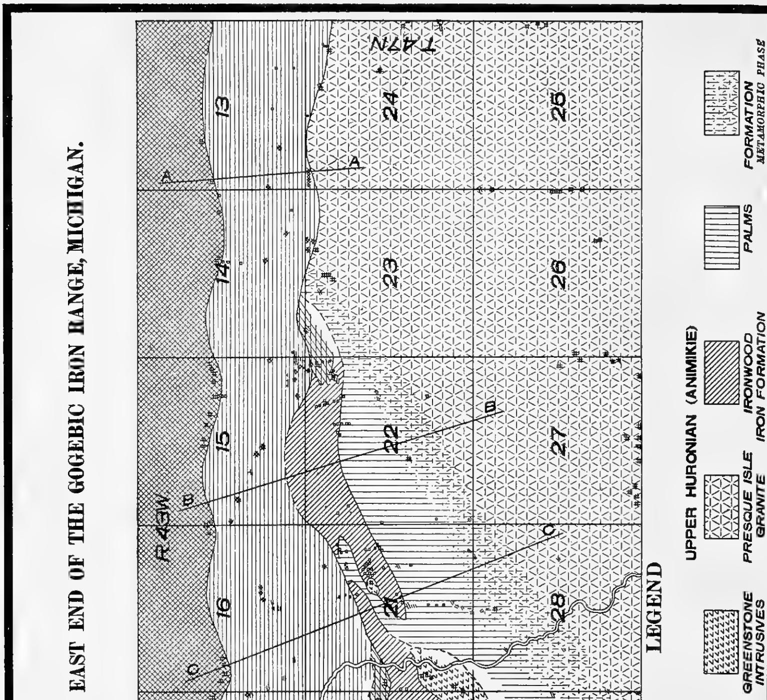

GEOLOGIC MAP OF A PART OF THE EAST END OF THE GOGEBIC IRON RANGE, MICHIGAN.
R 44 W.
R 43 W.
T 47 N
LEGEND
KEWEENAWAN
LAVAS, SANDSTONE AND CONGLOMERATE
POST-ANIMIKIE
COPPS FORMATION
GREENSTONE EXTRUSIVES
GREENSTONE INTRUSIVES
UPPER HURONIAN (ANIMIKIE)
PRESQUE ISLE GRANITE
IRONWOOD IRON FORMATION
PALMS
FORMATION METAMORPHIC PHASE

Fig. 1

units only to section 22, T. 47 N., R. 44 W., where a new unit representing the ferruginous slates is introduced to carry their combined stratigraphic horizons east to the extremity of the range, although isolated patches of iron formation are correlated with the Ironwood.

The writers' studies indicate very clearly: (1) the extension of the Palms and Ironwood formations eastward 7 miles beyond the point at which they are dropped as cartigraphic units from the

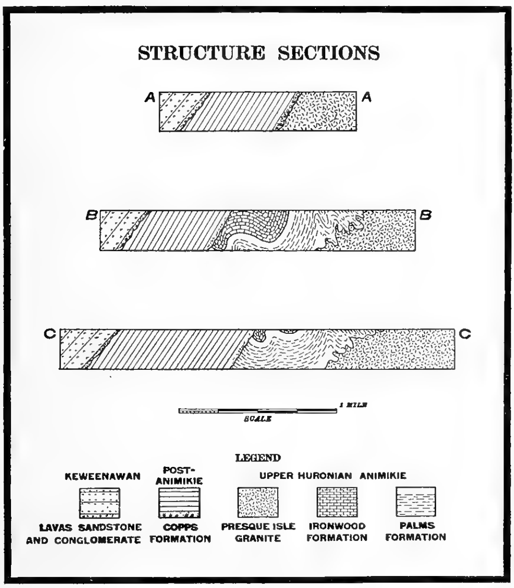


FIG. 2

maps of the United States Geological Survey; (2) these formations disappear, *not through lithologic gradation* into ferruginous slate, but through *abrupt truncation by a superior unconformable series;* and (3) that *these formations are intruded by the granite which appears on these maps as Archean* (Laurentian).

The Palms formation.—The Palms formation west of section 22, T. 47 N., R. 44 W., contains about 400 feet of graywacke and slate with interbedded thin bands of quartzite which become gradually more prominent in the upper horizons and grade into a

top layer of heavily bedded quartzite 50 feet thick. The basal horizon is a conglomerate from 1 to 10 feet thick. East of this township the massive upper quartzite is generally persistent, but the lower horizons grade irregularly into micaceous and hornblendic schists toward the contact with the intrusive Presque Isle granite in which, so far as we have determined, the base of the Palms is involved. The former correlation of the metamorphosed lower horizons of the Palms formation with the Keewatin is based on early field studies and has unfortunately been carried without revision into later work.

The Ironwood formation.—The iron-bearing member overlies abruptly but conformably the heavy upper quartzite of the Palms formation. A thin basal layer is in some places finely detrital.

The Tyler slate, which overlies the Ironwood formation west of Sunday Lake, is abruptly cut out at the great Wakefield overthrust and does not appear on the east end of the range. From Wakefield east to section 16, T. 47 N., R. 44 W., the Ironwood formation is overlain by the Keweenawan series, and beyond this locality with the Copps group, by which it is trunkated and cut out in the N.W. ½ of section 22, T. 47 N., R. 43 W.

Except where altered by intrusives, to which further reference will be made, the Ironwood formation is made up of ferruginous chert and slate, iron ore, cherty iron carbonate, and interbedded black and gray slate. In T. 47 N., R. 44 W., it is split, 200 feet above its base, by a great graywacke-slate member about 500 feet thick, which is identical in character with the slate forming the so-called "secondary foot wall" of the Brotherton, Sunday Lake, and Castile mines at Wakefield and is apparently at the same stratigraphic horizon. The inclusion of this slate member for a distance of not less than six miles east of Sunday Lake gives the Ironwood formation a maximum thickness of 1,300 to 1,400 feet.[1]

[1] Further observation west of Sunday Lake in both Michigan and Wisconsin indicates that the base of the Ironwood is conglomeratic throughout almost its entire extent. An intra-formational conglomerate and quartzite, first observed by geologists of the Oliver Iron Mining Company, occurs in the Pabst mine at Ironwood, a little less than 500 feet vertically above the Palms quartzite. A similar conglomerate has been observed at about the same horizon by W. O. Hotchkiss, state geologist of Wisconsin, in some of the mines on the Wisconsin end of the range. So far

Extrusive and intrusive greenstones.—Abundant exposures of greenstone occur throughout four square miles in T. 41 N., Rs. 43 and 44 W., and in a few other scattered small areas. These are mainly of extrusive types, but some coarsely crystalline intrusive equivalents are associated with them. Later dikes of diabase, probably of Keweenawan age, intrude the greenstone. The intrusives are mainly volcanic agglomerate and breccia, but in association with them are amygdaloidal lavas, tuffs, and some slate and conglomerate. An interesting feature of the lavas is a clearly indicated genetic relation to inclosed bands and irregular masses of red jasper. Basic flows with well-developed amygdaloidal tops are interbedded with the Ironwood formation from as far west as the Palms mine eastward to near Wakefield but are thus far not known to occur on the east end of the range.

The volcanics are closely associated with the Ironwood formation, in part interbedded in its upper horizons but in greater part superimposed. That they are unconformably overlain by the Copps group is an inference based on general field relations and the inclusion of pebbles of the volcanics in the Copps conglomerate rather than direct observation of contacts. Wherever the Copps group is in juxtaposition to the volcanic greenstones it invariably dips away from them.

The evidence indicates that volcanism was active on sea and land from middle Ironwood time, continuing perhaps for an indefinite period after the close of sedimentation.

The Presque Isle granite.—We propose the name Presque Isle granite for the rocks described by Van Hise and Irving as constituting the "Eastern Laurentian Granite Area."[1]

as observations have gone it may be said that this conglomerate bears no structural discordance to the underlying member of the Ironwood formation.

Mr. Fred Wolff has recently described the occurrence of a conglomerate in middle horizons of the Biwabik formation on the Mesabi iron range (*Engineering and Mining Journal*, C, No. 3). Inasmuch as the Ironwood and Biwabik formations are believed to be equivalent, this conglomerate may indicate a widespread disconformity. However, this disconformity is of very minor significance as compared with the profound unconformities which separate the Upper Huronian from the Animikie and the Animikie from the Lower Huronian, and, so far as known, has little importance in the consideration of the major subdivisions of the Huronian series.

[1] *Monograph 19, U.S. Geol. Survey.*

In a later publication[1] we will introduce evidence, based on field surveys and drilling operations, in support of the correlation of the Presque Isle granite with those occurring southwestward and southeastward to the state line in Michigan and over an adjacent large area in northern Wisconsin, possibly including the later intrusive granites of the Florence and Menominee districts, heretofore doubtfully correlated with the Keweenawan.[2] The close of Animikie time over this great area seems to have been characterized by batholithic intrusions of granite on a grand scale, comparable in the pre-Cambrian of the Lake Superior region to those of Laurentian time.

The Presque Isle granite is predominantly an acid type which varies through gradational phases into syenite and diorite. The rocks are mainly massive, but marked gneissosity is not an uncommon structural feature.

Relations to Upper Huronian (Animikie) group.—The inference that the Presque Isle granite is intrusive in the Animikie group rests on the following evidence:

1. The lower members of the Palms formation show a gradual transition in proportion as they approach the granite to micaceous and hornblendic schists and gneisses with obliteration of clastic structure.

2. The Palms and Ironwood formations are characterized by abundant quartz veins and the Palms formation also by veins of aplite and pegmatite which become increasingly prominent toward the granite.

3. The Ironwood formation becomes amphibolitic, magnetitic, and schistose in proportion to approach of the granite.

4. The Palms formation is apparently absent in the vicinity of section 25, T. 47 N., R. 44 W., where the Ironwood formation is an amphibole-magnetite schist and is separated by only a short distance from exposures of the granite.

The foregoing relations may be illustrated by reference to a few important localities. The best exposed section of the Palms formation

[1] "Contributions to the Pre-Cambrian Geology of Michigan and Northern Wisconsin," *Publication 18, Michigan Geol. and Biol. Survey* (in press).

[2] *Monograph 52, U.S. Geol. Survey.*

is adjacent to a north-south line from the center of section 21 southward about one mile to the granite near the center of section 28, T. 47 N., R. 43 W. The upper quartzite member underlies the Ironwood formation near the center of section 21. Proceeding southward there are numerous exposures and test pits showing the characteristic lower horizons of interbedded quartzite, quartz slate, and graywacke, complexly folded and cut by numerous quartz veins. From about 800 paces south the veins of aplite and pegmatite appear and the rocks become noticeably micaceous and hornblendic, with a gradual disappearance of clastic structure until they give place to those more properly described as mica-hornblende schist and gneiss. This mineralogical transition is accompanied by a gradual increase in the size and number of the aplitic and pegmatitic veins. The granite first occurs in intimate interbanding with hornblende gneiss and finally excludes it entirely.

In the vicinity of section 25, T. 47 N., R. 44 W., and section 30, T. 47 N., R. 43 W., the Presque Isle granite has apparently eaten its way upward through the upper heavy quartzite of the Palms into the Ironwood formation, and has very probably displaced or absorbed both of these formations in the E. $\frac{1}{2}$ of the former township. In these localities the Palms formation is apparently absent and the Ironwood formation is represented by amphibole-magnetite schist, which in section 25 is separated from hornblende syenite by a narrow strip about 150 feet wide in which there are no exposures.

In the workings of the Presque Isle mine on section 21, T. 47 N., R. 43 W., a conspicuous feature of the Ironwood formation is the occurrence of an unusual abundance of anastomosing quartz masses and veins. Pit dumps and drill records indicate that this feature is not local but general throughout the iron formation in this township. This phenomenon coupled with other evidence of granitic intrusion has considerable significance.

We take care to state that complete mineralogical gradation of either member of the Animikie into granite cannot be observed in any single exposure, but that such gradation is a fact and accounts for the absence of identifiable contacts is a conclusion which rests on field observations which admit of no other interpretation.

Copps group (*Post-Animikie*).—The Copps group is the name which we propose for the rocks occupying the area which is mapped and described in *Monograph 19, U.S. Geol. Survey,* under the heading "Ferruginous Slates of the Eastern Area." It is composed of a basal conglomerate, overlain by a great thickness of graywacke slate, which is highly ferruginous in the western half of its area and is associated in certain horizons, especially the lower ones, with considerable non-clastic chert and here and there with jasper.

The important member of this formation for purposes of this article is the basal conglomerate. This conglomerate is exposed in many places and shows a characteristic variation in content of matrix and pebbles with the character of the formation on which it rests.

From section 23, T. 47 N., R. 43 W., east to the extremity of the range it is in contact with the Presque Isle granite; the conglomerate here is composed of granite pebbles and bowlders ranging up to 4 or 5 feet in diameter imbedded in a matrix of sideritic arkose, the bowlder content exceeding the volume of matrix. A few pebbles of chert, jasper, quartzite, and greenstone appear in a large exposure near the S.W. $\frac{1}{4}$ of section 24, T. 47 N., R. 43 W., but one mile west of this place, near the old Copps mine, where the Copps formation truncates the Animikie, the granite pebbles have entirely disappeared, giving place to smaller fragments of chert and quartzite, chert predominating, derived from the Ironwood and Palms formations. The matrix here has increased in relative importance and has changed to finely clastic chert and quartz cemented by iron carbonate. North of the center of section 21 and near the N.W. corner of section 19, T. 47 N., R. 43 W., the conglomerate is exposed in test pits. In these localities the matrix has the character of graywacke in which are inclosed numerous small rounded pebbles of chert, jasper, and vein quartz. The conglomerate has also been penetrated by drilling in section 21, T. 47 N., R. 43 W., and, according to J. M. Louggeor, Jr., occurs "all along the north and east edges of the greenstone area" in sections 13, 14, 19, and 24, T. 47 N., R. 44 W.[1]

[1] A thesis on the geography of part of the Gogebic Range. Harvard University, 1915 (unpublished).

Relations of the Copps group to the Upper Huronian (Animikie).—It has been stated that the unconformity separating the Copps and Animikie groups is one of the first order of magnitude. This conclusion rests on the following evidence:

1. The bottom horizon of the Copps group is a true basal conglomerate.

2. The Presque Isle granite intrudes the Animikie but yields bowlders and matrix to the basal conglomerate of the Copps.

3. The Copps group is in great structural discordance with the Animikie.

4. The areal relationships are those of unconformity.

5. Volcanic rocks are characteristic of the Animikie and are not known to occur in the Copps formation.

The only one of these evidences remaining to be discussed refers to the great structural discordance between the Upper and Middle Huronian groups. At this point the reader is referred to the accompanying map and sections which exhibit the areal and structural relationships much more satisfactorily than any amount of written description could possibly do. It should be stated in this connection that the structural sections are based to a great extent on drill records, as are also the areal relationships of the Ironwood and Palms formations in sections 20, 21, and 22, T. 47 N., R. 43 W. It is not held that other interpretations of structure as far as concerns merely the details are not possible, but that those interpretations which have due regard to the facts of occurrence must coincide in major outline and cannot fail to establish the main fact of great structural discordance between the Upper and Middle Huronian groups. The data for structural sections are unusually ample because of the large amount of drilling and test-pitting which has been executed, fortunately in the most critical area, from the geologic point of view, on the east end of the range.

The dominant structure of the Animikie group of the Gogebic Range is that of a simple monocline in approximate structural parallelism with the Keweenawan series. Pre-Keweenawan deformation by folding is conspicuous only in the vicinity of Wakefield and on the east end of the range. Without venturing an opinion as to the vicinity of Wakefield, to which locality our field studies

have not extended, it is clear that the main folding of the Animikie on the east end is not only pre-Keweenawan but also pre-Copps. The cause and the localization of the deformation are referable to batholithic intrusion of granite.

Relations to Keweenawan series.—Studies of the Keweenawan series were limited to the problem of its relation to the Copps group.

The Keweenawan is unquestionably unconformably superimposed on the Copps. The contact is marked in some places by a strong basal conglomerate carrying pebbles and bowlders derived from all of the members of the Copps and Animikie groups. Inasmuch as this conglomerate has not been described, perhaps a brief reference to it will not be out of place here. In sections 14, 15, and 16, T. 47 N., R. 43 W., the base of the Keweenawan is a coarse, firmly cemented, sandstone 50 to 75 feet thick grading into conglomerate near the base. The pebbles of the conglomerate comprise fragments of vein quartz, red slate, quartzite, graywacke, jasper, chert, and iron ore.

Relations to Cambrian (Eastern or Jacobsville) sandstone.—The Eastern or Jacobsville sandstone has been described by other writers as resting unconformably on rocks here designated as the Copps group. This evidence rests on an examination many years ago of exposures in test pits. These pits are now filled and overgrown with brush and we were therefore unable to add anything to existing information.

RELATIVE TO THE BEARING OF CONCLUSIONS ON LAKE SUPERIOR PRE-CAMBRIAN CORRELATIONS

It is not our purpose to anticipate in this paper the following discussion of the bearing of this revision of the geology of the Gogebic Range on the more general problem of the pre-Cambrian correlations. To the student of Lake Superior geology the suggestion will arise that had these relationships been discerned thirty years ago the history of the pre-Cambrian correlations would probably have been markedly different, not only in Michigan but in general.

Any disposition which may be made of the Copps group in the correlations must take into account the magnitude of the uncon-

formity at its base and the bearing on the general problem of the enormous scale of the batholithic granitic intrusions of late Animikie time.

DISCUSSION OF CORRELATION

C. K. LEITH AND R. C. ALLEN

INTRODUCTORY STATEMENT

The non-productive eastern end of the Gogebic Range is folded, faulted, and associated both with intrusive and extrusive rocks. In the lack of sufficient exposures and exploration scarcely more than lithologic mapping was possible when the range was worked over by Irving and Van Hise.[1] Structural mapping was not attempted. In recent years, exploration of this area by drilling has connected some of the isolated outcrops of iron-formation bands and has made it possible to map the area structurally. This has been done effectively by Messrs. Allen and Barrett, who outline their results in the preceding paper. While there are still some parts of the area to which the structure and areal connections are obscure, certain features of exceptional interest from the standpoint of general Lake Superior geology have already been worked out. These are: (1) discovery of the fact that there is an intrusive granite with important metamorphic effects in the supposed eastward continuation of the Gogebic iron formation (Animikie); and (2) the existence of a ferruginous graywacke formation with jasper and chert bands, called the Copps formation, separated from the inferior Animikie series by a great structural unconformity and unconformably below the Keweenawan.

These facts raise interesting questions as to revision of the pre-Cambrian correlation of Michigan, which touch more or less also the correlations of the Lake Superior region as a whole. Leith thinks it possible that the Copps series may be of only local significance and require no general change, but he agrees with Allen that Michigan correlation should be carefully tested to ascertain whether the known facts can be better expressed by a different

[1] *Monograph 19, U.S. Geol. Survey*, 1892.

correlation than that heretofore used. Two main possibilities present themselves: (1) that the Copps represents a fourth Huronian series between the Animikie and Keweenawan; (2) that there are three Huronian series, and that the Copps formation may be equivalent to certain series elsewhere in Michigan, now otherwise correlated.

1. CORRELATION ON THE ASSUMPTION THAT THE COPPS FORMATION REPRESENTS A FOURTH HURONIAN SERIES BETWEEN THE ANIMIKIE AND KEWEENAWAN

If the Copps series be regarded as the remnant of a much larger series that has everywhere else been removed by erosion, it may be necessary to give it a position of the same general order as that of the Animikie or Upper Huronian, with the result that the Huronian series of Lake Superior become relatively lowered throughout the column, which would involve some changes in names. There would then be four Huronian series instead of three. The evidence, as it is, seems to afford too slight a basis for moving downward all the rest of the formations and introducing sweeping changes in nomenclature. The names now used have come to have well-understood significance, and well express the principal similarities and relationships of the pre-Cambrian series. When the extent and relationships of the Copps series can be more fully demonstrated it may be desirable to introduce such sweeping changes. To do so now would be premature and would almost inevitably require further changes when the significance of the Copps formation becomes more fully understood.

2. CORRELATION ON THE ASSUMPTION THAT THERE ARE THREE HURONIAN SERIES AND THAT THE COPPS FORMATION MAY BE EQUIVALENT TO CERTAIN SERIES NOW OTHERWISE CORRELATED

Allen and Barrett regard the Copps formation as the equivalent of the Upper Marquette series, the Princeton series,[1] and certain formations in other districts above the Animikie which are believed to be, as in the Gogebic district, below the Keweenawan. Leith

[1] R. C. Allen, "Correlation and Structure of the Pre-Cambrian Formations of the Gwinn Iron-bearing District of Michigan," *Jour. Geol.*, XXII (1914), 560–73.

believes that the evidence thus far available, while suggestive, does not constitute a sufficient basis for throwing over the present correlation, which brings out the main features of succession in common in the different districts, and that a revision introduces difficulties in local correlation quite as great as any in the old correlation. In order to bring out clearly the possibilities in the situation, the two views are presented below respectively by R. C. Allen and C. K. Leith.

A REVISION OF THE CORRELATION OF THE HURONIAN GROUP OF MICHIGAN AND THE LAKE SUPERIOR REGION

R. C. ALLEN

The Huronian group comprises at least three unconformable series of pre-Cambrian sedimentary and associated igneous rocks separated from the younger Keweenawan series by a great unconformity and from the older Archean system by another more profound unconformity. The position of the group has been defined by the United States Geological Survey and by a committee of Canadian and United States geologists[1] (1904) as follows:

Paleozoic		
		Unconformity
	Keweenawan	
		Unconformity
		Upper
		Unconformity
Algonkian	Huronian	Middle
		Unconformity
		Lower
		Unconformity
	Keewatin	
Archean		Eruptive contact
	Laurentian	

Former correlations of the Huronian group as well as the entire pre-Cambrian of the various districts of the Lake Superior region in the United States are mainly the results of detailed field study and mapping by the United States Geological Survey. These

[1] *Jour. Geol.*, XIII (1905), 89–104.

studies have been published in a series of monographs on the most important districts and a final summary covering the entire region.[1]

In 1892 Irving and Van Hise published their completed work on the Gogebic iron range. They believed that the Huronian series in the Gogebic district comprises *two* unconformable series, the Upper and the Lower Huronian, and that the Upper Huronian or iron-bearing series is probably equivalent to the Animikie of the north shore of Lake Superior. In 1896 Van Hise, Bayley, and Smyth issued a monograph on the Marquette iron range. They found here *two* unconformable Huronian series, the *Upper* and the *Lower* Huronian, which were correlated with the *Upper* and the *Lower* Huronian of the Gogebic Range, the Negaunee iron-bearing series of the Marquette Range falling in the lower Huronian. These two works are the origin of the *dual classification* of the Huronian which was gradually extended to cover all of the other districts of the Lake Superior region until A. E. Seaman discovered, about 1902, that the "lower" Huronian of the Marquette Range is divisible by a great unconformity at the base of the Ajibik quartzite. In 1904 Seaman's discovery was formally recognized by an international committee of Canadian and United States geologists which adopted the tripartite classification of the Huronian group.[2] The Negaunee iron-bearing series was separated from the Lower Huronian to form the new Middle Huronian, but the old correlations outside the Marquette Range were preserved. In 1913 Allen accounted for a tripartite division of the Huronian in the Gwinn synclinorium and correlated the Gwinn iron-bearing series with the Negaunee

[1] R. D. Irving and C. R. Van Hise, "The Penoke Iron Bearing Series of Michigan and Wisconsin," *Monograph 19, U.S. Geol. Survey*, 1892; C. R. Van Hise and W. S. Bayley, "The Marquette Iron Bearing Series of Michigan, with a chapter on the Republic Trough, by H. L. Smyth," *Monograph 28, U.S. Geol. Survey*, 1896; C. R. Van Hise, J. M. Clements, W. S. Bayley, and H. L. Smyth, "The Crystal Falls Iron Bearing District of Michigan," *Monograph 36, U.S. Geol. Survey*, 1899; C. K. Leith, "The Mesabi Iron Bearing District of Minnesota," *Monograph 43, U.S. Geol. Survey*, 1903; J. M. Clements, "The Vermilion Iron Bearing District of Minnesota," *Monograph 45, U.S. Geol. Survey*, 1903; W. S. Bayley, "The Menominee Iron Bearing District of Michigan," *Monograph 46, U.S. Geol. Survey*, 1904; C. R. Van Hise and C. K. Leith, "Geology of the Lake Superior Region," *Monograph 52, U.S. Geol. Survey*, 1911.

[2] *Jour. Geol.*, XXII, No. 6 (1914).

iron-bearing series of the Marquette district. In 1914 Allen and Barrett found that the Upper Huronian, as described by Van Hise and Irving, on the east end of the Gogebic Range includes *two* unconformable series separated by a profound unconformity. This discovery has, in their judgment, opened the way for a revision of the correlation of the Huronian group of the Lake Superior region in which some of the difficulties and inconsistencies in the old classification largely disappear.

The new correlation places the Animikie series in the Middle Huronian, rather than in the Upper Huronian as in former correlations. Briefly stated, the steps in the argument are these:

1. The Animikie (iron-bearing) series of the Gogebic Range is the equivalent of the Negaunee (iron-bearing) series which constitutes the Middle Huronian of the Marquette Range; therefore, the Animikie series of the other Michigan districts, Minnesota, and the north shore of Lake Superior is also the equivalent of the Negaunee series, i.e., Middle Huronian.

2. The Negaunee series of the Marquette Range is unconformably overlain by the Upper Huronian. The Ironwood (Animikie) series of the Gogebic Range is also overlain unconformably by a series which is *unconformably beneath the Keewenawan.* This series is equivalent to the Upper Huronian of the Marquette Range. The Animikie is therefore Middle Huronian.

CORRELATION OF THE ANIMIKIE SERIES AS MIDDLE HURONIAN

Correlation of the Ironwood series with the Negaunee series.—The Ironwood (Animikie) iron-bearing series of the Gogebic Range is correlated with the Negaunee (Middle Huronian) iron-bearing series of the Marquette Range because (1) these series occupy identical positions in the Huronian succession of these districts, (2) are essentially similar, (3) are underlain and overlain by essentially similar series, (4) are in practically adjacent territory, and (5) there is substantial evidence of their equivalence through direct connection of the Negaunee series with the Vulcan (iron-bearing) series of the Crystal Falls–Iron River district which bears the same relation to a great granite batholith and its outliers as does the

Ironwood series of the Gogebic Range and its correlatives in the Marenisco and Turtle ranges.

The similarity of the successions in the Gogebic and Marquette districts is striking, and coupled with proximity would ordinarily determine a direct correlation of the similar series occupying identical positions in the group even were there no further evidence of identity.

CORRELATION TABLE III

Correlation of the Huronian Group in the Gogebic and Marquette Districts

	Marquette District	Gogebic District
	Greenstone intrusives and extrusives	
	Clarksburg volcanics partly replacing Michigamme slate	Graywacke and slate
Upper Huronian	Michigamme slate carrying iron-bearing lenses (Bijiki schist)	Ferruginous and cherty slates and jasper
	Goodrich conglomerate—quartzite	Conglomerate
	Unconformity	Unconformity
	(Deep erosion)	Presque Isle granite
		Tyler slate
Middle Huronian (Animikie)	Negaunee (iron-bearing) formation and extrusive greenstone	Ironwood (iron-bearing) formation and extrusive greenstone
	Siamo slate	Palms quartzite and quartz slate
	Ajibik quartzite	
	Unconformity	Unconformity
	Wewe slate	(Deep erosion)
Lower Huronian	Kona dolomite	Bad River dolomite
	Mesnard quartzite	Sunday quartzite
	Unconformity	Unconformity
Archean		

The absence of slate above the Bad River dolomite should be considered with the evidence of deep erosion in Middle-Lower Huronian time in the Gogebic district which not only removed the slate if it was ever present there but also the entire Lower Huronian over the greater part of the range. In respect to the Middle

an

	Vermilion	Gunflint Lake	Animikie	Felch
M				Mica sch Ferrugin caceou.
c in- rma-	Rove slate Gunflint iron formation	Rove slate Gunflint iron formation	Black slate Iron formation	Vulcan i tion Felch sch
	Granite, dolorites, lamprophyres intrusive into rocks below	Greenstone and granite intrusives	Granite and greenstone intrusive into rocks below	Randvill Sturgeon
lom-	Knike Lake slate	Graywacke		
	Agawa formation Ogishke conglomerate		Slate, graywacke, and conglomerate	

CORRELATION TABLE II

(Showing changes in Huronian correlation on the basis that the Animikie Group is Middle Huronian)

	Marquette	Gwinn	Sturgeon	Felch Mountain	Calumet	Menominee	Florence
Upper Huronian	*Greenstone intrusives and extrusives* *Michigamme slate partly replaced by Clarksburg volcanics* *Bijiki schist (iron bearing)* *Goodrich quartzite*	*Michigamme slate* *Goodrich graywacke and conglomerate*		*Mica sch[illegible] and mi-* *Ferrugin[illegible] [illegible]artzite*			*Quartzite and conglomerate*
Middle Huronian (Animikie)	*Negaunee iron formation* *Siamo slate* *Ajibik quartzite*	*Slate* *Negaunee iron formation* *Slate* *Arkose conglomerate*	*Iron formation (Negaunee)*	*Vulcan [illegible] forma-tion* *Felch schist*	*Hanbury slate* *Vulcan iron formation* *Felch schist*	*Granite* *Quinnesec schist* *Hanbury slate* *Vulcan iron formation* *Quartzite*	*Granite* *Quinnesec schist* *Hanbury slate* *Vulcan iron formation*
Lower Huronian	*Wewe slate* *Kona dolomite* *Mesnard quartzite*		*Randville dolomite* *Sturgeon quartzite*	*Randvill[illegible] [illegible]olomite* *Sturgeon [illegible]artzite*	*Randville dolomite* *Sturgeon quartzite*	*Quartzite* *Randville dolomite* *Sturgeon quartzite*	
Laurentian							
Archean Keewatin							

	Crystal Falls	Iron River	Vieux Desert Conover	Manitowish	Turtle	Marenisco
Upper Huronian	*Michigamme slate*	*Michigamme slate*				
Middle Huronian (Animikie)	*Paint slate?* *Greenstone intrusives and extrusives* *Hanbury slate* *Vulcan-Negaunee iron formation* *Ajibik quartzite* *Mansfield slate* *Hemlock volcanics*	*Paint slate?* *Greenstone intrusives and extrusives* *Hanbury slate* *Vulcan iron formation* *Slate*	*Granite* *Slate and schist*	*Granite* *Kyanitic, biotitic, garnetiferous, and graphitic schists* *Amphibole-magnetite rocks*	*Granite* *Greenstone intrusives and extrusives* *Slate* *Iron formation* *Slate* *Quartzite*	*Presque Isle granite* *Greenstone extrusives and intrusives* *Slate* *Iron formation* *Graywacke and quartzite*
Lower Huronian		*Dolomite and quartzite (Saunders formation)*			*Dolomite* *Dolomitic quartzite*	
Laurentian						
Archean Keewatin						

	Gogebic	Cuyuna	Mesabi	Vermilion	Gunflint Lake	Animikie
Upper Huronian	*Copps formation*					
Middle Huronian (Animikie)	*Presque Isle granite* *Tyler slate* *Greenstone extrusives* *Ironwood iron formation* *Palms formation*	*Basic and acidic intrusives and extrusives* *Virginia (St. Louis) slate* *Deerwood iron formation* *Slate* *Quartzite*	*Embarrass granite(?)* *Acidic and basic intrusives* *Virginia slate* *Biwabik iron formation* *Pokegama quartzite*	*Rove slate* *Gunflint iron formation*	*Rove slate* *Gunflint iron formation*	*Black slate* *Iron formation*
Lower Huronian	*Bad River limestone* *Sunday quartzite*		*Giants Range granite* *Graywacke conglomerate*	*Granite, dolerites, lamprophyres intrusive into rocks below* *Knife Lake slate* *Agawa formation* *Ogishke conglomerate*	*Greenstone and granite intrusives* *Graywacke*	*Granite and greenstone intrusive into rocks below* *Slate, graywacke, and conglomerate*
Laurentian						
Archean Keewatin						

Huronian (Animikie) of these districts the situation is reversed, i.e., the thick Tyler slate formation above the Ironwood series is to be considered with the evidence of deep erosion of the Negaunee series and the development of iron ores on the exposed surface of the iron formation (now represented by the hard ores of the upper part of the Negaunee formation) prior to the deposition of the Upper Huronian. If the correlative of the Tyler slate was ever deposited on the Negaunee formation, it had been removed prior to the deposition of the Goodrich quartzite. The Middle Huronian (Animikie) of both districts is characterized by volcanic activity which continued on into and culminated in the Upper Huronian in the Marquette Range but terminated prior to the deposition of the Copps formation in the Gogebic Range.

There is a strong resemblance of the Copps formation of the Gogebic to the Upper Huronian of the Marquette Range in respect to lithology and order of succession. At the base of these series is a great conglomerate which is overlain by slate and graywacke. Neither series carries a great productive iron-bearing member, but both contain jasper and ferruginous beds near the base, which in the Marquette district are locally iron ore-bearing.

It can hardly be doubted that if Van Hise and Irving had discovered in 1892 the great unconformity at the base of the Copps formation on the Gogebic Range, and Van Hise and Bayley in 1896 the great unconformity at the base of the Ajibik quartzite on the Marquette Range, the correlation of the Huronian group of the Lake Superior region would have been *tripartite* from the beginning and not *dual*, for the correlations have been built up on these two type districts which were earliest studied by the United States Geological Survey. The Ironwood (middle) series would *then* have been correlated with the Negaunee (middle) series because, as above stated, they occupy identical positions in the Huronian succession, are essentially similar, are overlain and underlain by essentially similar series, and are in practically adjacent territory. The character of the reasoning which was employed in correlating the formations in these districts and which resulted in a separation of the similar iron-bearing series in the correlation would have united them if the facts of present information

had been then available. There could *then* have been little if any reason for assigning these two iron-bearing series to different positions in the Huronian group and there is even less reason for such assignment today, because there is substantial evidence for the correlation of the Ironwood and the Negaunee series on other grounds, which will be discussed below.

CORRELATION OF THE VULCAN (IRON-BEARING) SERIES WITH THE NEGAUNEE (IRON-BEARING) SERIES

Marquette and Crystal Falls districts.—The basis of correlation of the formations of the Marquette and Crystal Falls districts is afforded by an indicated practical continuity of the Negaunee and Vulcan iron-bearing formations. In 1903 H. L. Smyth traced the Negaunee iron formation from the Republic trough southwest around two major anticlines to the northeast side of the great oval anticline in the northern part of the Crystal Falls district.[1] The relation of the Negaunee formation in this area to a persistent magnetic line may be seen on Fig. 3. From the vicinity of Michigamme Mountain, T. 44 N., R. 31 W., where the Negaunee formation is exposed, a magnetic line extends north through the Sholdice and Doan explorations, where the Negaunee is again exposed, and thence, north, northwest, west, and southwest around the great oval anticline into section 27, T. 46 N., R. 33 W., where it still coincides with the position of the iron formation. A short distance beyond the latter locality the line is broken, but it reappears after an interval of about 2 miles and passes through the Red Rock and Hemlock mines at Amasa and beyond, connecting with the iron formation which has been mapped as Vulcan or Upper Huronian.

The United States Geological Survey in 1911[2] accepted the conclusion that the magnetic line, exposures, explorations, and drift bowlders prove the practical continuity of the Negaunee formation from Michigamme Mountain for a distance of 25 miles around the great anticline to a point about a mile south of section 27, T. 46 N., R. 33 W., but from a point about 2 miles farther on,

[1] See Smyth's discussion in *Monograph 36, U.S. Geol. Survey*, pp. 452–55.

[2] *Monograph 52.*

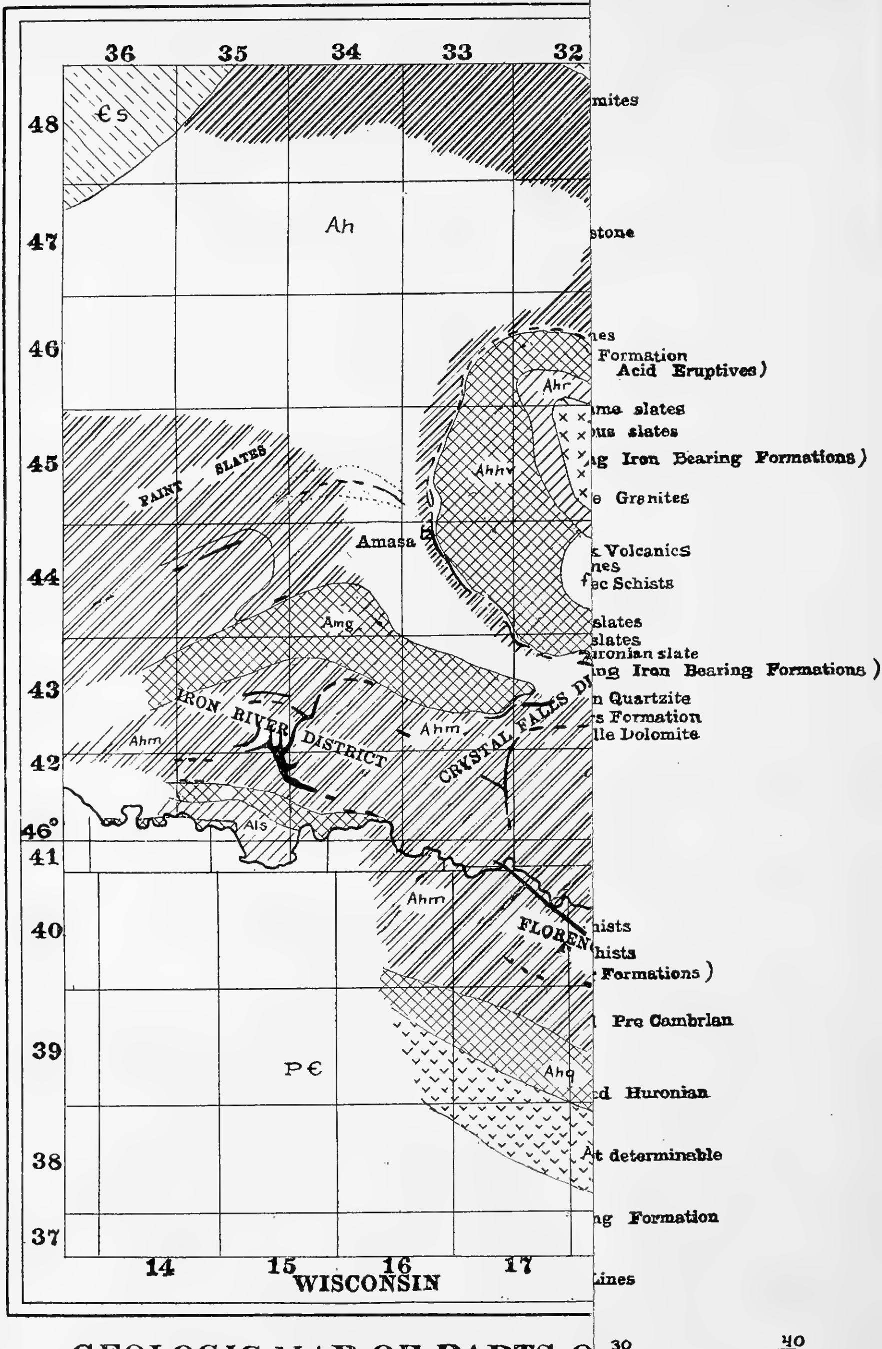
through the Red Rock mine and southward, the iron formation is correlated with the Vulcan or Upper Huronian notwithstanding the facts (1) of similar position with reference to the underlying Hemlock volcanics, (2) that the strike of the magnetic line north of the Red Rock mine indicates a continuity of the Vulcan iron formation there with the Negaunee a little farther north, and (3) that there is no evidence whatever beyond some difference in degree of metamorphism to show that the iron formations in section 27, T. 46 N., R. 33 W., and at the Red Rock mine are not one and the same.

The reason why the iron-bearing series (Vulcan) of the Crystal Falls–Iron River–Florence district was not correlated with the Negaunee by the United States Geölogical Survey is a simple one. In the Menominee Range, Van Hise and Bayley found that the iron-bearing series is unconformably above a quartzite-dolomite succession similar to the Lower Huronian of the Marquette Range and, under the dual classification, therefore Upper Huronian. Since the Negaunee formation (prior to 1904) had been correlated with the *Lower* Huronian, the question arose as to the position of the iron-bearing series in the intervening districts. It was reasoned that (1) because the iron-bearing series of the Crystal Falls district was at that time inseparable from the great Upper Huronian slate area opening out south and west from the Marquette district, and (2) because the Hanbury slate of the iron-bearing series of the Menominee Range seems to have areal connections with the slates of the Florence–Crystal Falls district, therefore the iron-bearing series in the latter district must be considered Upper Huronian. This conclusion was preferred, notwithstanding the evidence of continuity of the Vulcan and Negaunee formations in the northern part of the Crystal Falls district, and, as we shall see, became a source of considerable difficulty in applying the dual classification to the facts of succession in the Felch Mountain and Florence districts.

Recent developments in the Hemlock and Michigan mines at Amasa have fortunately determined conclusively that the Vulcan iron formation there which had been correlated as Upper Huronian is in reality Negaunee or Middle Huronian. On the thirteenth

level of the Hemlock and Michigan mines the folds in the iron formation are truncated by a heavy conglomerate and quartzite carrying fragments of the Negaunee formation of all sizes up to several feet in diameter, including small, angular, hard jasper fragments, rounded pebbles of chert and ore, and great bowlders of the iron formation. It is reported that this same conglomerate was found by drilling in similar relations to the iron formation about three miles south of Amasa and in section 36, T. 44 N., R. 33 W., about four miles farther southward. There can be no reasonable doubt that this conglomerate-quartzite is the Goodrich formation of the Marquette Range, where exactly similar relations are observed.

The productive iron formation at the Hemlock mine extends with only a few unexplored breaks in drift-covered country southeastward around the great anticline of Hemlock volcanics and other rocks and is believed to be almost if not quite continuous with the iron formation passing through the Hollister, Armenia, and other mines in the vicinity of Crystal Falls. Such continuity is indicated, so far as definite information is available, by drilling, underground openings, and magnetic surveys. Furthermore, the iron formation on the west and southwest sides of the great Crystal Falls oval anticline maintains the same position with reference to the underlying Hemlock volcanics that it does on the north and east sides. Therefore if the iron formation at the Red Rock and Hemlock mines is Negaunee, the burden of proof rests on those who would assert in the absence of any supporting facts that the Upper-Middle Huronian unconformity cuts out the Negaunee iron formation and occupies an inferior position with reference to the Vulcan iron formation at any or all points southward. This practical continuity was accepted by Clements in 1899 and by Leith and Van Hise in 1911, as shown on the maps issued in *Monographs 36* and *52* of the United States Geological Survey. In fact, Leith and Van Hise argue in this work that the iron formation at the Hemlock mine is *not* Negaunee *on the assumption of practical continuity with the Vulcan formation of the Crystal Falls district to the south and lack of continuity with the Negaunee formation a few miles northward.*



GEOLOGIC MAP OF PARTS O

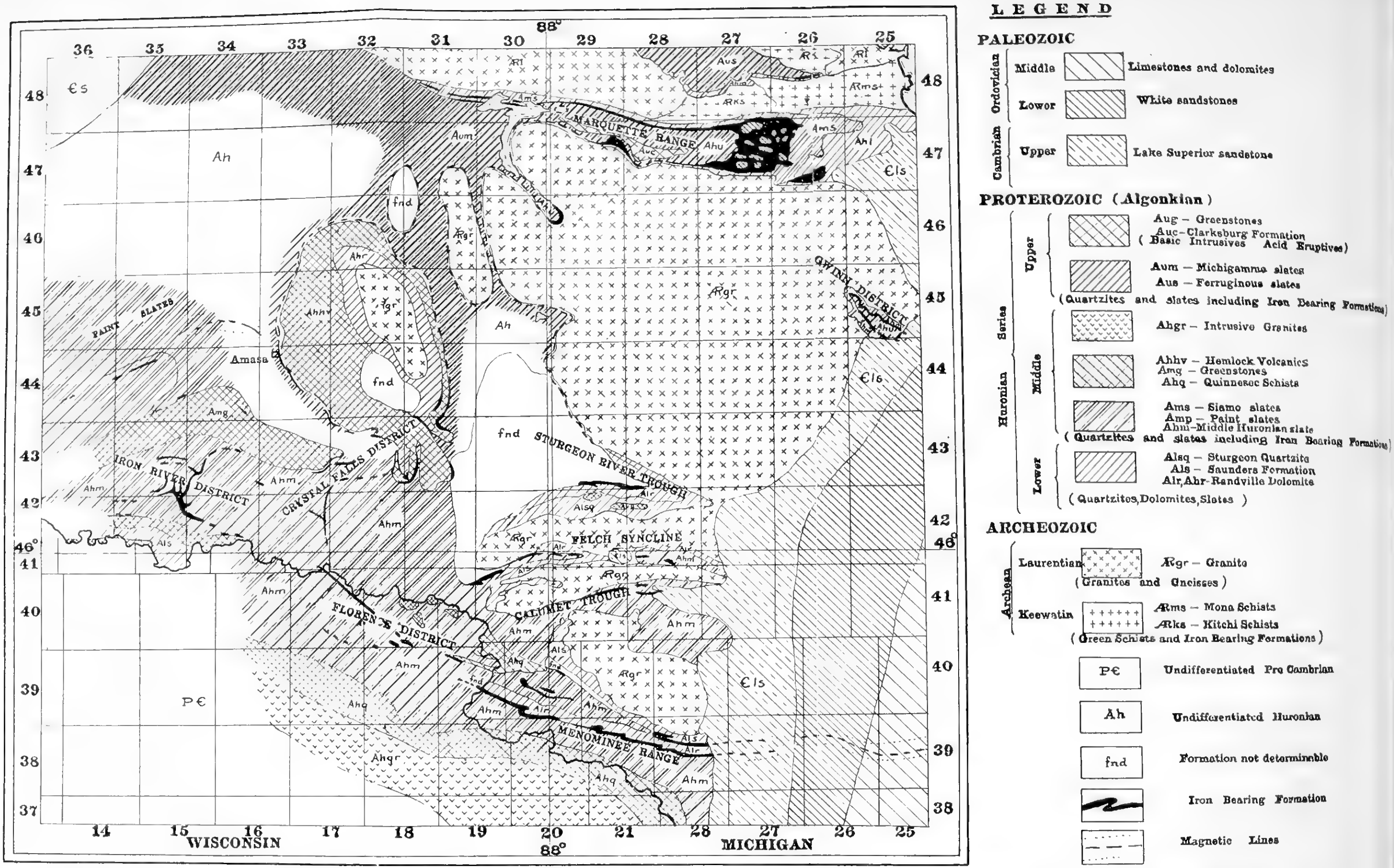
CORRELATION OF THE NEGAUNEE (IRON-BEARING) SERIES WITH THE VULCAN (IRON-BEARING) SERIES OF THE STURGEON TROUGH, FELCH MOUNTAIN DISTRICT, CALUMET TROUGH, AND MENOMINEE RANGE

These ranges lie south of the Marquette district and east of the Iron River–Crystal Falls–Florence district and form eastward-projecting tongues of the Huronian of this great area (see Fig. 3).

1. *Sturgeon River syncline.*—The Negaunee formation has been traced by outcrops, exploration, and magnetic surveys from the Marquette Range into the north limb of the Sturgeon River syncline. From near Witch Lake, about 8 miles south of Republic, the Negaunee formation is shown by the mapping of the United States Geological Survey to rest directly on the Archean. In the Sturgeon syncline, however, the Archean is overlain by the Randville dolomite and Sturgeon quartzite, equivalent to the Kona and Mesnard of the Lower Huronian in the Marquette Range. On the south limb of this syncline the iron formation reappears above the Randville dolomite and has naturally been correlated with the Negaunee of the north limb.

2. *Felch Mountain district.*—The Felch Mountain district is a narrow syncline of Huronian rocks downfolded in the Archean, from one to two miles wide, trending east-west, and, like the Sturgeon trough, opening out westward into the great slate area of the Crystal Falls district, wherein the structure is obscured. It is separated from the Sturgeon trough north of it by an anticline on which the Archean appears as a belt of granite about $2\frac{1}{2}$ to 3 miles wide. In the Felch syncline there is an iron formation (Groveland) similar to that in the Sturgeon trough, separated from the quartzite-dolomite below by conformably underlying sedimentary schist (Felch schist) which has not been observed in the Sturgeon trough, although it may be present there also. Bayley makes no mention of the Negaunee formation of the Sturgeon trough in 1899,[1] although it was subsequently discovered through exploration and is shown on the maps of the United States Geological Survey published in 1911.[2] On these maps the iron formation in the Sturgeon trough is called Negaunee (Middle Huronian) and in the Felch syncline

[1] *Monograph 36, U.S. Geol. Survey.* [2] *Monograph 52.*



GEOLOGIC MAP OF PARTS OF MICHIGAN AND WISCONSIN

10 0 10 20 30 40
SCALE MILES

FIG. 3

Vulcan (Upper Huronian). This seems to be an arbitrary classification and represents the necessity of making the change somewhere from the Negaunee of the Sturgeon trough to the Vulcan of the Menominee Range. Lithologic similarity, proximity, and, so far as known, similarity of succession were cast aside in favor of the supposition that the Felch schist opens out and connects with sediments to the west which had been correlated as Upper Huronian, although the area in which the connection is indicated is deeply drift-covered and devoid of rock exposures. But even if this connection were a fact, as it may be, it constitutes in our opinion merely an added reason why the Groveland iron formation should be correlated with the Negaunee, since the Crystal Falls slate-iron formation series has been shown to be more probably *Middle* rather than Upper Huronian.

As a matter of fact Smyth *did* correlate the Groveland with the Negaunee formation in 1899, but after Seaman's discovery of the unconformity at the base of the Ajibik quartzite necessitated the correlation of the Negaunee series with the *Middle* Huronian (1904), Van Hise and Leith in 1911 took the Groveland out of the *Lower* Huronian and placed it in the *Upper* Huronian. In order to make this change it was necessary to assign a highly metamorphic quartzite-mica schist series which is unconformably *above* the Groveland to the Keweenawan or Paleozoic, for the reason that no place was then left for it in the Huronian group. Smyth was obviously right in correlating the Groveland with the Negaunee formation. The quartzite-mica schist series above the Groveland bears no resemblance to the Keweenawan or Paleozoic. It is Huronian, and we believe it should be correlated with the Copps, part of the Michigamme[1] and the Princeton series. The quartzite which is unconformably above the slate-iron formation series (Animikie) of the Florence district is similarly correlated.

3. *Calumet trough.*—In the Calumet trough, about 4 miles south of the Felch syncline, the situation is practically identical with that in the Felch Mountain district, and the same arguments for revision of the correlation apply here as in the Felch district. In other words, the iron formation now assigned to the Upper

[1] For division of the Michigamme series, see discussion below.

Huronian is believed to be really an equivalent to the Negaunee of the Middle Huronian. Rocks similar to the quartzite-mica schist series of the Felch syncline are exposed at the old Hancock exploration in T. 41 N., R. 27 W., and in at least one or two places near the southern edge of the Calumet trough. Each of the three Huronian series seems to be represented here exactly as in the Felch syncline a few miles north.

4. *Menominee Range.*—The general similarity of the iron-bearing series of the Menominee district with that in the Sturgeon, Felch, and Calumet troughs and in turn with certain phases of the Negaunee formation of the Marquette district, its similarity in relation to the underlying Lower Huronian, and its areal connections with the great slate-iron formation series of the Florence–Crystal Falls district determine that the Vulcan of the Menominee Range is probably of Negaunee age, and that if the iron formation in the districts intermediate between the Menominee and the Marquette ranges is Negaunee there is no basis on lithological, structural, or other grounds for assigning the iron formation of the Menominee Range to any horizon other than Negaunee, i.e., Middle Huronian.

Leith has described a remnant of cherty quartzite, of a maximum thickness of 70 feet, lying apparently unconformably between the Randville dolomite below and the Traders member of the Upper Menominee series above in the vicinity of Norway,[1] but he places no emphasis on it so far as concerns its significance in the correlations. This formation, according to the more recent opinion of Dr. Leith,[2] is a remnant of regolith unremoved by erosion in Lower-Middle Huronian time.

CORRELATION OF THE IRONWOOD SERIES (ANIMIKIE) OF THE GOGEBIC RANGE WITH THE VULCAN (MIDDLE HURONIAN) SERIES OF THE CRYSTAL FALLS–IRON RIVER–FLORENCE–MENOMINEE DISTRICT ON THE BASIS OF SIMILAR RELATIONS TO INTRUSIVE GRANITE

Area southeast of the Gogebic Range including Marenisco, Turtle, Manitowish, Vieux Desert, Conover, Iron River, Crystal Falls, and Menominee districts.—Probably the most striking feature of the

[1] *Monograph 52, U.S. Geol. Survey,* pp. 234–35.

[2] As expressed in conversation with the writer.

Middle Huronian (Animikie) of these districts is the general prevalence of intrusive granite. Heretofore these granites have been variously correlated, from Laurentian in the Gogebic, northern Wisconsin, and Menominee districts through the Upper Huronian in the Crystal Falls district to Keweenawan in the Florence-Menominee and northeastern Wisconsin areas.

In the Menominee district Bayley found that the granite south of the Menominee River intrudes a series of basic volcanics called the Quinnesec schist, which he correlated with the Keewatin. Although it was realized that the correlation of the Quinnesec schist as Keewatin introduced a conception of structure quite out of accord with natural inferences on the basis of the facts, it remained for Corey and Bowen, working under the direction of Van Hise and Leith in 1905,[1] and Hotchkiss in 1910,[2] to show conclusively that the Quinnesec schist is partly intrusive into, but in greater part interbedded with, the upper part of the Upper Huronian, i.e., Animikie.[3]

Inasmuch as the granite was thus proven to be the youngest rock in these districts, and the youngest pre-Cambrian sediments had been correlated with the *Upper* Huronian, Leith and Van Hise in 1911[4] correlated the granite with the Keweenawan and extended the boundaries south and east to include several thousand square miles of acid intrusives in north-central Wisconsin which had been mapped and described by Weidman in 1905.[5]

It is interesting to note here that Brooks and Wright had correctly interpreted these relations as early as 1876, although they did not, at that time at least, comprehend the extent of the great mass of intrusive granite. To quote from Brooks:

> In the summer of 1874 Chas. E. Wright and myself, exploring the country west and south of the Menominee River about 90 miles from its mouth, under the auspices of the Wisconsin Geological Survey, observed a large granite

[1] Unpublished notes of field work done in 1905 by G. W. Corey and C. F. Bowen.

[2] Unpublished field notes of W. O. Hotchkiss.

[3] Middle Huronian of this article.

[4] *Monograph 52, U.S. Geol. Survey.*

[5] S. Weidman, "The Geology of North Central Wisconsin," *Bull. Wis. Geol. and Nat. Hist. Survey,* No. 16, 1907.

area, the north edge of which was bounded by dark colored hornblendic and micaceous schists of Huronian age, which I have since concluded are equivalents of the youngest member of that series yet observed in the Marquette iron region. . . . . The lithologic character of this wide granite belt bore so much resemblance to the Laurentian rocks, which are extensively developed on the waters of the Sturgeon River in Michigan, 10 to 20 miles to the northeast, that we were disposed at the time to believe that some phenomena of folding or faulting had brought rocks belonging to that system to the surface in an unexpected quarter. Professor Pumpelly and myself, several years previously, had observed, farther to the north and west, similar granitic rocks crossing the Michigamme and Paint rivers (branches of the Menominee), presenting similar puzzling relations with beds known to be Huronian, and younger than as well as lithologically different from any rock then known to be of that period.

A careful consideration of all of the facts to be observed in the Menominee region confirms me in this hypothesis. . . . .[1]

The conclusion of Brooks is confirmed by the work of Hotchkiss in the Florence district of Wisconsin in 1910.[2]

In 1911–14 the writer and assistants, through field mapping and diamond drilling, traced what appears to be this same granite from the Iron River district westward through the Animikie series of the Vieux Desert–Conover district and Manitowish, Turtle, and Marenisco ranges, in all of which it is in intrusive relation with the sediments, and connected it with the "Eastern Laurentian area" of Van Hise and Irving on the eastern Gogebic Range, which was considered by these geologists to form a part of the Archean basement complex, whereas it actually intrudes the Lower and Middle Huronian series and is unconformably overlain only by the Upper Huronian (Copps) series. *Thus the great granite batholith of northern Wisconsin has been fairly demonstrated to be, not only in intrusive relations with the Animikie sediments over several thousands of square miles, but also to be overlain unconformably by a pre-Cambrian series which is unconformably below the Keweenawan.*

Inasmuch as the Vulcan (Animikie) series of the Crystal Falls, Iron River, Menominee, Florence, and other Michigan districts has been shown to be very probably equivalent to the Negaunee

[1] T. B. Brooks, "On the Youngest Huronian Rocks South of Lake Superior and the Age of the Copper Bearing Series," *Am. Jour. Sci.*, Vol. II (1876), 206–7.

[2] Unpublished manuscript.

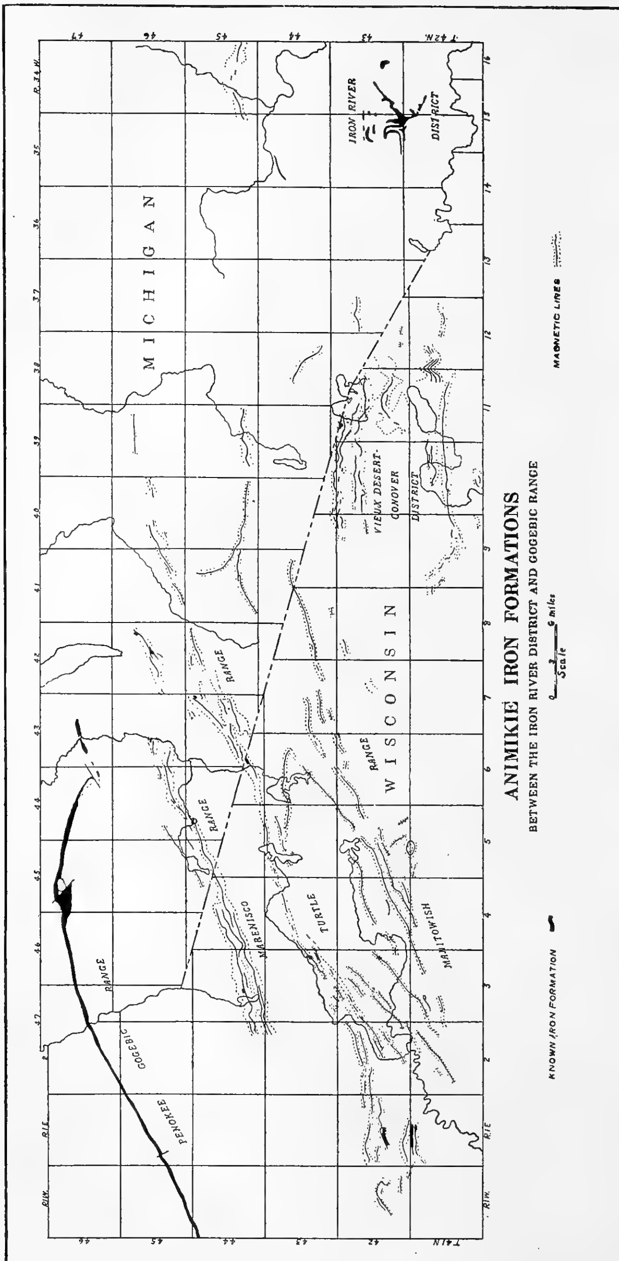

ANIMIKIE IRON FORMATIONS
BETWEEN THE IRON RIVER DISTRICT AND GOGEBIC RANGE
MICHIGAN
WISCONSIN
IRON RIVER
DISTRICT
VIEUX DESERT-
CONOVER
DISTRICT
RANGE
TURTLE
MANITOWISH
MARENISCO
RANGE
GOGEBIC
RANGE
PENOKEE
KNOWN IRON FORMATION
MAGNETIC LINES
Scale
6 miles

FIG. 4

series of the Marquette Range and the Ironwood series of the Gogebic Range, the age of this great granite batholith and its outliers may be considered as late Middle Huronian.

CONSIDERATION OF ARGUMENTS WHICH MAY BE ADVANCED FOR THE RETENTION OF THE CORRELATION OF THE ANIMIKIE AS UPPER HURONIAN

We have seen that the correlation of the Animikie as Upper Huronian dates from the earliest work of the United States Geological Survey in the Lake Superior region. It seems reasonable to believe that if the early workers in the Gogebic and Marquette ranges had recognized that the Huronian of both exhibits a *tripartite* rather than a dual division, the Animikie would have been considered by them as Middle Huronian rather than Upper Huronian. We have referred to some of the difficulties which were encountered in the application of the dual classification to some of the other Michigan ranges and have shown how these difficulties largely disappear in the correlation of the Animikie series as Middle Huronian. Having established the basis of the proposed correlation, we may turn to a consideration of some of the objections which may be offered to it.

Dr. Leith has emphasized the general similarity of the slate in the great Iron River–Crystal Falls area and the Animikie slates of the Cuyuna and Mesabi districts of Minnesota, and believes that "there is probably but one great slate formation of *this type* in the Lake Superior region." Although this observation has little bearing on the question of whether the Animikie is Middle or is Upper Huronian, it may be observed that this similarity is marked so far as concerns the slates which occur in areal association with the iron formation, but does not extend to the great mass of slate which occupies a large area of indefinite boundaries north and west of the Crystal Falls–Iron River iron-bearing area. These rocks are mainly graywacke of uniform composition and appearance and are not at all similar to the fine-grained gray and black pelites of the Animikie series which are associated with the iron formation (Vulcan) in the Iron River, Crystal Falls, Florence, and Menominee districts and the Negaunee iron formation of the Marquette Range.

The contrast is so marked that in 1910 I described these graywackes under the distinct formation name[1] of Paint slate and called attention to the structural evidence indicating that they are stratigraphically *above* the Vulcan (Animikie) series.

The correlation of the Paint formation presents a difficult problem. It constitutes a large part of the Michigamme series heretofore correlated with the Upper Huronian. We have seen that there is strong evidence that the Vulcan slate-iron formation series of the Iron River–Crystal Falls district, which has been described as a part of the Michigamme, is Middle Huronian. That part of the Michigamme slate which is *known* to be Upper Huronian is confined to a limited area adjacent to the western part of the Marquette iron range and the northern part of the Crystal Falls district where it is plainly unconformably above the Negaunee iron-bearing series. If the Paint slate is *conformable* with the Negaunee-Vulcan, it follows that the part of the Michigamme slate which is Upper Huronian must rest unconformably on it, and from the areal distribution of these formations it appears that the plane of the unconformity should intersect the erosion surface somewhere *north* of the Paint formation north of the Iron River district. Unfortunately, a general scarcity of exposures in the area where the unconformity may be expected to occur and the difficulty of identifying horizons over any considerable distance in formations of these kinds will render the discovery and mapping of this unconformity, if it exists, difficult. It is being searched for at the present time.

As bearing on the probable age of the Paint formation it may be said here that the Wolf Lake granite, north of Watersmeet in T. 46 N., R. 39 W., is intrusive in quartz-mica schist which, judging from lithology and areal relationships, is a metamorphic equivalent of the Paint slate. The Wolf Lake granite is correlated with the Presque Isle granite of the Gogebic Range and that in the intervening territory and is believed to be an outlier of the great granite batholith of northern Wisconsin. The Presque Isle granite intrudes the *Middle* Huronian (Animikie) and yields detritus to the Upper

[1] R. C. Allen, "The Iron River Iron Bearing District of Michigan," *Publ. 3, Mich. Geol. and Biol. Survey*, 1910.

Huronian. If the Presque Isle and Wolf Lake granites are of the same age, it follows that the Paint formation is Middle Huronian.[1]

In the argument for retention of present correlations which follows, Dr. Leith is unable to accept the correlation of the Vulcan series of the Crystal Falls–Iron River district with the Negaunee series of the Marquette Range, for the reason, among others, that the Vulcan iron formation occurs in lenses in slate, as against the well-defined position of the Negaunee formation in the stratigraphic succession. In this connection it should be observed that this dissimilarity of occurrence is fully as great and of the same order in comparison with the Animikie series of the Gogebic, Mesabi, and old Menominee ranges with which the Vulcan is correlated by him as well as by me. If the observation is of importance in the one case, is it not equally so in the others?

In further reference to the correlation of the Negaunee and Vulcan iron-bearing series, it may be observed that the Negaunee series in the Marquette trough is thick and well developed and should be expected to occur in the down folds over a large area. The Ajibik quartzite has a maximum thickness of 700–750 feet, the Siamo slate probably from 500–600 feet, and the Negaunee iron formation in excess of 1,000 feet. The Siamo slate is not persistent on the south side of the trough west of the latitude of Ishpeming, where the Negaunee rests directly on the Ajibik quartzite. The Negaunee formation, as well as the underlying quartzite, becomes thinner in this direction, and in the Sturgeon trough and around the great anticline in the Crystal Falls district the Ajibik formation, according to the latest maps of the United States Geological Survey (1911), is entirely absent, the Negaunee formation there resting directly on the Lower Huronian. The correlation of the Vulcan formation of the Crystal Falls district with the Negaunee formation of the Marquette district is criticized by Dr. Leith on the basis of the apparent absence of the Ajibik quartzite in the former area. However, *the Ajibik formation is known to have disappeared from beneath the Negaunee in the direction of the Crystal*

[1] R. C. Allen and L. P. Barrett, "Contributions to the Pre-Cambrian Geology of Michigan and Northern Wisconsin," *Publ. 18, Mich. Geol. and Biol. Survey* (in press).

Falls district, as indicated on the latest maps of this district published by the United States Geological Survey.

As to the importance which may be attached to the absence, in the southern part of the Crystal Falls district, of the formations which underlie the Ajibik quartzite adjacent to the iron formation in the Marquette Range, it may be observed that these formations appear not only on the great anticline in the northern part of the Crystal Falls district but on the upturned limbs of the synclines to the east—the "eastward-projecting tongues of this great area," wherein erosion has cut down into the Archean.

Finally, it may be said that it is easier to account for the disappearance of a coarse clastic basal member which is obviously related to a shore phase of deposition than to justify, in the absence of definite supporting facts, the assumption of a rather abrupt disappearance of an entire series.

It may be objected further to the correlation of the Vulcan and Negaunee formations that the hard ore horizon of the upper part of the Negaunee is not duplicated in the Vulcan of the Crystal Falls district. This fact has little bearing on the correlation because this phase of the Negaunee is related, as shown by Van Hise and others, to a controlling structural condition which nowhere occurs in the Crystal Falls district, i.e., it is the result of deformational-metamorphic alteration adjacent to the heavy competent Goodrich quartzite-conglomerate. This relationship, areally considered, is entirely fortuitous and may be expected to appear only where the structural conditions and deformation on which it is dependent are in evidence. So far as *lithology* is concerned the Vulcan and Negaunee formations are in other respects practically identical.

Dr. Leith opens his discussion of the Upper-Middle Huronian unconformity at the Hemlock mine and vicinity as follows: "Practically no new evidence has been brought in, in the crucial area between the Marquette and Crystal Falls districts, and such additional evidence as there is, is damaging to the conception of the new hypothesis." The discovery of that unconformity has apparently reversed Dr. Leith's ideas concerning the correlation of the iron formation in the northern part of the Crystal Falls district which he recently (1911) correlated with the Vulcan but

now correlates with the Negaunee and refers to it as "distinctively Negaunee type." If the Hemlock iron formation and the Vulcan were considered the same in 1911 and prior years, what *new evidence* may be cited on which they may be separated *now?* The only *new evidence* is the discovery of the fact that the iron formation at the Hemlock mine and for several miles north and south of it is Negaunee, and this is therefore precisely the evidence to which Dr. Leith refers as damaging to the new hypothesis and in confirmation of the old correlation.

Dr. Leith has also laid emphasis on the fact that the unconformity at the Hemlock mine has not been traced across the northern part of the Iron River district and therefore that proof is lacking that any part of the Michigamme slate occupies an inferior position with reference to this unconformity. This is a weakness in the new hypothesis, but it is certainly not greater than the weakness in the argument which has been advanced that this unconformity cuts out the Negaunee formation in the northern part of the Crystal Falls district and in some unknown place passes beneath the Vulcan formation of the southern part of the district. If the greenstone belt north of the Iron River district is Hemlock (Middle Huronian) as correlated by Clements as well as by Van Hise and Leith (1911), there is an anticlinal structure here which *demands* that the unconformity turn "back on itself at some place south of Amasa" and carry "out westward to the north of the Iron River–Crystal Falls district" (see Fig. 3). This would seem a necessary consequence of Dr. Leith's published interpretation of the structure with which I agree (in reversal of my opinion in 1909) because of the discovery of the unconformity in question and the evidence based on the recent drilling in the vicinity of Crystal Falls. In brief, the structural facts point to a closely compressed syncline between the two masses of volcanics, similar to those northeastward including the Republic trough, with the formations involved sharply upturned on its opposite limbs.

GENERAL REMARKS ON THE CORRELATION OF THE ANIMIKIE SERIES WITH THE MIDDLE HURONIAN

The correlation of the Animikie series with the Middle Huronian eliminates what would otherwise be the necessity of assuming a

fourth Huronian series of which the Copps formation of the Gogebic Range would be the sole representative. The practical identity of the Huronian succession of the Gogebic and Marquette ranges, together with the marked similarity of the Copps formation and the Upper Huronian of the Marquette and Gwinn districts, is believed to be an adequate basis for moving the Animikie of the Michigan and other Huronian districts downward into the Middle Huronian, particularly as this correlation, as we have seen, eliminates what would otherwise be a further necessity of including the uppermost series of metamorphic sediments in the Felch, Calumet, and Florence districts in the Keweenawan or the Paleozoic, where they obviously do not belong.

The anomalous position in the correlations of the great Negaunee iron-bearing series has been unsatisfactory to many students of the pre-Cambrian for many years.[1] We now have a firm basis for the correlation of the Negaunee series with the great productive iron-bearing series of the Animikie of all of the other districts of the Lake Superior region and are able to recognize the consideration that the unique conditions which resulted in the deposition of the great Huronian iron formations were regional rather than local and should be correlated in time. While it is true that the names now used have come to have well-understood significance because of long usage, it cannot be held that this is a valid argument for the retention of a classification which no longer fairly interprets the facts of present knowledge.

ARGUMENTS FOR RETAINING THE PRESENT CORRELATION

C. K. LEITH

Doubts as to the present correlation of the Michigan formations center in the great slate area (Michigamme slate) carrying the iron formation of the Iron River, Crystal Falls, Florence, and con-

[1] See A. C. Lane and A. E. Seaman, *Mich. Geol. Survey*, *Annual Report*, 1908, pp. 23–30. In presenting their mature conclusions on Michigan pre-Cambrian sucessions, Lane and Seamen correlate the Animikie iron-bearing series throughout with Negaunee (Middle Huronian) of the Marquette Range.

tiguous districts, which lies between the well-determined succession of the Gogebic on the one hand and the well-determined succession of the Marquette on the other. The close folding of this area, its intrusion by granite, its relatively soft character, causing low relief and few exposures, the fact that exploration and development in this area are much less advanced than in the Gogebic and Marquette districts, the fact that there is a general absence of distinctive quartzite or other formations as horizon-markers, have made it difficult to carry correlations with certainty through this great slate area. In general, the slates show evidence of shallow water or delta deposition in their rapidly alternating bedding, current bedding, ripple marks, graphitic layers, and considerable thickness. In all these respects it differs from any of the slates, other than Animikie, in the pre-Cambrian of Lake Superior. It is substantially like the slate above the Animikie iron formation of the Gogebic and Mesabi districts, like the Animikie of the Cuyuna and St. Louis River districts, and like the Michigamme slate of the Upper Marquette series (Animikie). Throughout the great slate areas of this type in the Lake Superior region the iron formation has somewhat similar characteristics, being in irregular folded, more or less discontinuous, lens-like bands within the slates, without definitely determinable horizon. The occurrence and nature of the ores are so much alike throughout these great slate areas that they are naturally grouped by the explorers or mining men.

Nearly all investigators of the geology have noted these broad similarities and have correlated the slates of these districts as Animikie.

In arguing for the revision of pre-Cambrian correlation of Michigan, Allen accepts the correlation of the slate series of the Iron River–Crystal Falls district with the Animikie or Upper Huronian of the Gogebic district, but its correlation with the Upper Huronian or Animikie of the Marquette district is disputed on the ground that it may be the equivalent to the Middle Huronian of the Marquette district, leaving the Upper Huronian of the Marquette district available for correlation with the newly discovered Copps series. The discovery of this new series leads, naturally, to a search for a more extensive occurrence in the region, but, in my

judgment, the proposed change runs squarely across certain fundamental facts which will not give way to the new hypothesis, however attractive it may be. Allen's argument is based (1) on the supposed equivalence of the Iron River slates (and iron formation) with the Animikie in the Gogebic district, (2) on the supposed equivalence of the Iron River slates with the Middle Huronian of the Marquette district.

1. That the slates and iron formation of the Iron River district are to be correlated with the Animikie of the Gogebic district is the conclusion that has been reached by nearly all investigators, and in the absence of definite proof to the contrary must stand. I would suggest, however, that the recent discovery of the Copps series opens up an alternative, that the Iron River series is equivalent to the new Copps series and both equivalent to the Upper Huronian of the Marquette district. This would have the effect of throwing the Animikie down a step in the scale, but would not disturb the local correlation between the Marquette and Crystal Falls–Iron River districts. This is only a possibility, for outcrops are few in the connecting areas, metamorphism due to intrusion has been intense, and I see no likelihood for some years of sufficient information being available to develop the evidence necessary for this possible conclusion.

2. That the slates and iron formation of the Iron River–Crystal Falls district are equivalent to the Negaunee (Middle Huronian) of the Marquette district is a possibility which has been often considered discarded. Practically no new evidence has been brought up in the crucial area between the Iron River–Crystal Falls district and the Marquette district, and such additional evidence as there is, is damaging to the conception of the new hypothesis.

The great essential fact of the situation, which cannot be ignored, is that in the Marquette district the Negaunee iron formation (Middle Huronian) is unconformably below a great slate series (Michigamme slates, Upper Huronian) which contains lenses of iron formation identically like those in the Crystal Falls and Iron River districts, and contrasting with the Negaunee iron formation of the Negaunee. The Michigamme slate of the Marquette district is all but proved to be the same as that in the Iron River and Crystal

Falls region. They are areally contiguous, contain the same lithologic types, and throughout contain iron formation of remarkably uniform type and occurrence. To assume that they are not of the same age requires not only the introduction of an entirely hypothetical unconformity, but also the assumption that the Negaunee formation has entirely changed its character to the exact extent necessary to make it similar to the slate-iron formation of the Crystal Falls–Iron River district. Such a thing is of course possible, but extremely improbable.

Nowhere in the Crystal Falls–Iron River district has the close folding and erosion disclosed the distinctive Siamo slates of the underlying Ajibik quartzite characteristic of the horizons below the Negaunee. Neither has it disclosed any formations underlying the Ajibik close at hand, with the exception of certain volcanics. Somewhere they should appear on the erosion surface if the series were really Middle Huronian.

That there are two slates in the Iron River–Crystal Falls district, as argued by Allen, is possible, but no trace whatever of an unconformity has been found, and the differences in lithology which he cites as evidence are completely duplicated by differences in lithology in the Upper Huronian slates of the Marquette district.

The principal argument for a change in correlation is based on a supposed direct areal connection of the Negaunee iron formation of the Marquette district with the iron formation of the Iron River–Crystal Falls district. With some interruption, due to folding, the Negaunee iron formation has been traced southwestward from the Marquette district to a point in the northeast portion of the Crystal Falls district near Amasa, on the west side of an area of underlying Hemlock volcanics. Somewhere south of this area the character of the iron formation changes, for before reaching the iron formation belts in the slates (Upper Huronian) in the vicinity of Crystal Falls, the formation is of different character, the underlying fragmental phases are absent, and both walls are conformable slate. The difficulty of drawing a line here, in the absence of exposures and complete exploration, has long been obvious, and yet the contrast in types and occurrence of iron formation seem to require that some line be drawn. In successive maps the line has shifted more or less.

The general accuracy of the conclusion that the two formations are of different age has been strengthened by the finding of a basal conglomerate resting directly on the Negaunee formation and locally cutting it out in the general vicinity of Amasa. So far as this conglomerate has been traced (6 or 8 miles) it shows conclusively that the slate series on the west is unconformably above the Negaunee formation on the east. Where this unconformity goes to the south is not yet discovered, and hence there is still doubt as to the exact location of the line of contact between the Negaunee iron formation and the Upper Huronian. The extension of the unconformity along its trend south or southeast in general separates the formation of distinctly Negaunee type on the east from the formation of distinctly Iron River–Crystal Falls type associated with slates on the west. Allen, on the other hand, would turn the unconformity back on itself at some place south of Amasa and carry it out westward to the north of the Iron River–Crystal Falls district, in order to leave the Iron River–Crystal Falls iron formation with the Negaunee formation below the unconformity. It seems possible that the unconformity to the south of Amasa cuts out the Negaunee formation, allowing the Upper or Animikie series to lap over against the underlying formation, exactly as happens in places in the Marquette district. The statement that the known Negaunee formation is areally connected with the known Crystal Falls–Iron River iron formation is based on general similarity of trend of the iron formation, as shown by explorations and various magnetic belts. As the two have been folded together, this similarity of trend is to be expected in any case. There is plenty of room for the unconformity to run diagonally across this general trend at almost any point. Exactly the same argument for connection in the Marquette district would result in mapping Negaunee iron formation with Upper Huronian iron formation, whereas the two are really separated by a profound unconformity.

The writer is not impressed with the argument that if a threefold division of the Huronian had been first made, subsequent correlation would have been of a different sort. It is entirely true that different correlations might result in starting from a twofold or threefold basis of division, but the history of correlation of pre-

Cambrian formations in the Lake Superior country discloses no backwardness in revising correlations as fast as facts are found to warrant revision. Seaman's discovery of a tripartite division in the Marquette district was immediately recognized by the geologists of the United States Geological Survey and was first published by C. K. Leith in 1904,[1] with permission of Professor Seaman. Neither may it be said that there is any delay in recognizing the significance of the supposed new formation in the eastern Gogebic district. In fact, the writer took some part in the field in the analysis of the situation leading up to this discovery. It may be freely conceded that in the past there was more emphasis on a given number of series as a basis of correlation, but the basis of correlation has been constantly widened by the addition of new criteria. It seems peculiarly inappropriate, therefore, to argue that what might have been done in the past on the basis of a preconceived notion of a number of series should serve as a primary basis of classification now when a much greater variety of facts is available.

In short, the suggested changes in correlation seem to the writer interesting possibilities, for which evidence must be carefully looked for in further geological work in this district, but that they are only possibilities and in the present state of knowledge cannot stand against the general considerations above outlined. It should not be overlooked that the new series suggesting these sweeping changes is yet known in but a few outcrops in a limited area where the folding and intrusion have been extensive, and that there is distinct possibility that the formation may have been of only local significance.

In case the newly proposed classification can be established it will be a welcome advance in our knowledge of Lake Superior geology. The purpose of this argument is not to discourage an attempt to make such an advance, but to indicate the difficulties inherent in the problem and the impossibility in the present state of knowledge of accepting an interesting hypothesis as a proved fact.

[1] C. K. Leith, *Trans. Am. Inst. Min. Engrs.*, Lake Superior meeting, 1904.

PRESSURE AS A FACTOR IN THE FORMATION OF ROCKS AND MINERALS

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It is not surprising that application of the knowledge comprised in the older chemistry should have yielded comparatively little information as to the way in which rocks and minerals actually form; and for this reason. The older chemistry dealt in the main with the formation and behavior of substances at temperatures confined to a very small range and practically at a single pressure only—that of the atmosphere; in other words, it dealt with the merest slice of the surface which would represent the behavior of the substance throughout the range of temperature and pressure within which it is capable of existence. At the same time there was a failure to recognize that its work was so limited in character, and consequently the endeavor to deduce regularities from the behavior of substances at ordinary temperature and pressure—under arbitrary conditions, in other words—was but partially successful. The reason for this is clear if we consider how our ordinary chemistry—that of the 20° C. level—would be changed at the 200° level, for a large number of compounds would then have become unstable; and again, that the number of compounds persisting at the 1,000° level, and a fortiori at the 2,000° level, would be but a small fraction of those which exist at ordinary temperatures.

Igneous rocks and minerals have formed under conditions much removed from 20° and 1 atm. pressure; so that we are little likely to ascertain much about their formation except by thorough and extensive investigations over a wide range of temperatures and pressures, investigations which would at the same time throw much needed light on a number of very important chemical problems. Incidentally it may be observed that the variables temperature and pressure are completely analogous; that besides the ordinary freezing-point of a liquid—in stating which we imply a definite

pressure of 1 atm.—there is also the freezing-pressure at a definite temperature; for example, mercury freezes at -38.85° under 1 atm.; at 0° C. its freezing-pressure—the pressure which will cause it to freeze—is about 7,600 atm.

The purpose of the present paper is to discuss briefly the available experimental evidence in the light of the principles involved, to indicate the conclusions which the somewhat scanty observations seem to justify, and to point out the limitations to which such conclusions are necessarily subject. In general it would seem that the importance to geology of the effects of pressure upon changes such as the melting of a single pure substance—those, namely, usually regarded as physical changes—has been overestimated relatively to that of the influence of pressure upon systems containing more than one component—namely, upon chemical changes. The influence of pressure upon chemical equilibrium is especially marked if one or more of the components of the system are volatile, and must therefore be taken into account in any discussion of the behavior of the magma; indeed, the order of separation from a complex solution containing a volatile component depends just as much on whether, and how, the pressure changes as upon the mode and rate of cooling.

There has been very little direct experiment on the influence of pressure upon the formation of rocks and minerals, by reason of the technical difficulties in the way of making such experiments; so that we are forced in large part to reason from analogy with those substances whose behavior it has been possible to investigate. In reasoning from analogy we must, of course, bear in mind the inherent limitations to which conclusions thus derived are subject. On the other hand, in reasoning backward from present-day field evidence, one must also proceed with caution; for, presuming even that one is aware of all the factors in the net result as we see it, it is hard to disentangle the effects of these several factors, especially since the intensity of each is in general unknown. For example, we can make only a very rough guess at the temperature and pressure prevailing at the time when a given process was taking place, and consequently cannot gauge satisfactorily the relative importance of these factors in producing the result as we see it. Moreover, the

composition of a rock mass as we see it now is no absolute criterion of what was the total composition of the magmatic solution from which it separated; for we are ignorant of the character and amount of the volatile components which were present in the original liquid magma.

In discussing metamorphic processes—or indeed in any discussion of the effects producible by pressure—one must remember to distinguish carefully between uniform (or pure hydrostatic) pressure and non-uniform compression, which is equivalent to a stress; for the former is incompetent to produce some of the results which may be brought about by the latter. This distinction is important; failure to observe it has led to many apparently contradictory statements, and consequently to some confusion. The main difference is that whereas a homogeneous solid body recovers its original dimensions and properties when relieved from a uniform pressure, it is changed permanently by a stress, provided that the stress exceeds the elastic limit of the solid; this residual effect is evident as a change of shape—a deformation—which is accompanied by a change in several other properties, e.g., in the case of metals by changes in density, electric resistance, and thermoelectric power. That the effects of stress on a crystalline aggregate outweigh those of uniform pressure follows from thermodynamical principles if we identify a deformation under stress with a real, partial, local temporary melting, on the basis that the stress acts upon the solid but not upon the liquid phase.[1] This hypothesis, which is not in conflict with any direct experimental evidence, serves to correlate a number of observations on the flow of rocks and of metals, and has in recent years been utilized by several English metallographers as a means of accounting for phenomena connected with the important question of the hardening of metals. In this connection it may be pointed out that the circumstance that a solid has been deformed indicates, not that its strength was necessarily small, but that the shearing forces to which it was exposed were great enough to overcome the resistance to deformation offered by the solid; in other

[1] This hypothesis, and some of its consequences, have been discussed in previous papers; see Johnston and Adams, *Am. Jour. Sci.*, XXXV (1913), 206; Johnston and Niggli, *Jour. Geol.*, XXI (1913), 599.

words, the fact that a rock has flowed is no criterion of its strength, nor does it indicate that at the time of flow the rock was necessarily liquid as a whole.

Let us now consider briefly the effects of uniform pressure upon some of the properties of a pure crystalline substance, taking up first its melting-curve. For a pure crystalline substance there is, at a definite pressure, a single temperature—the melting-point—at which its solid and liquid forms can coexist indefinitely in equilibrium, this being the temperature at which the vapor pressure of solid and that of the liquid become identical. The melting-point is influenced by change of uniform pressure—such as we may conceive to be exerted by either a gas or a mobile liquid (e.g., oil) which is insoluble in the substance. By joining up the melting-points at various pressures we obtain the melting-curve, the slope of which (dT/dP) at any point can easily be shown to be

$$\frac{dT}{dP}=\frac{T\Delta V}{\Delta H}$$

where T, ΔV, and ΔH are respectively the melting-point, the change of volume, and the heat change (latent heat of fusion) at that particular point on the curve. From this equation it is obvious that the general shape of the curve is determined by the way in which ΔV and ΔH vary with the pressure; that the condition for a maximum melting-point is that ΔV should vanish while ΔH remains finite, and for a critical end-point solid-liquid that ΔV and ΔH should vanish simultaneously. But we cannot predict from theoretical principles how ΔH and ΔV will vary—in other words, the shape of the curve can be determined only by experiment. In this connection it may be noted that there is no general relation between ΔH and ΔV, for we may have a large latent heat associated with a small volume change and vice versa; indeed, in two particular cases we have a negative volume change with a positive heat effect, these two exceptional cases—ice I and bismuth—being the only pure substances whose melting-point is known to be lowered by pressure. There is little doubt that the silicates are normal in this respect, showing a rise of melting-point with increase of uniform pressure, because everything indicates that they expand on melting

—that at the melting-point the volume of the liquid is greater than that of the solid.

The best and most extensive experimental work is that of Bridgman,[1] who determined the melting-curve of a number of substances at pressures up to 12,000 atm.—a pressure corresponding to a depth below the surface of the earth of about 30 miles. Some of his data are presented in Table I, which shows especially the gradual diminution of the effect with each successive pressure increment. Bridgman also made direct measurements of the change of volume on melting (ΔV) throughout his pressure range, and found that it decreases slowly and at a continuously decreasing rate; the shape of the curve suggests that the change of volume would not become zero at any finite pressure. Moreover, by combining his data and using the foregoing equation he calculated ΔH; it does not tend toward zero at higher pressures, but remains approximately constant, showing if anything a tendency to increase with the pressure. The general conclusion from this work accordingly is that, up to pressures of 12,000 atm. at least, there is no indication of a maximum melting-point, still less of a critical end-point solid-liquid; on the contrary everything indicates that such points, if indeed they exist at all, can occur only at pressures altogether outside the range of possible experiment at the present time.

Table I illustrates another interesting point: namely, that at a pressure of about 5,800 atm. the melting-point of solid CO_2 has reached 31°, the critical temperature of the liquid; at pressures higher than this it sublimes or passes directly into a gas. Consequently it is in this case actually possible to pass directly from a *gas* (not a vapor, for it is above the critical temperature) to crystalline solid merely by change of pressure. This would indicate that application of sufficient pressure would convert even a gas into a solid, and suggests that the existence of any large body of gas deep in the earth is as little likely as the existence of much liquid.

All of this evidence then tends to corroborate the belief that the central portion of the earth is substantially solid, though this does not imply that this solid will not flow whenever opportunity

[1] P. W. Bridgman, *Proc. Am. Acad.*, XLVII (1911–12), 347; *ibid.*, XLVII (1911–12), 439; *Physic. Rev.*, III (1914), 126, 153.

occurs for it to do so, since the stresses to which it is exposed must be far in excess of its elastic limit, even at moderate depths.

TABLE I

MELTING-CURVES OF VARIOUS SUBSTANCES (AFTER BRIDGMAN)

PRESSURE KG. PER SQ. CM.*	MERCURY		POTASSIUM		SODIUM		NITROBENZOL		CARBON DIOXIDE	
P	*t*	Δ*t*	*t*	Δ*t*	*t*	Δ*t*	*t*	Δ*t*	*t*	Δ*t*
1	−38.85		62.5		97.6		5.6		−56.6	
		5.10		16.2		8.6		21.9		19.3
1,000	−33.75		78.7		106.2		27.5		−37.3	
		5.09		13.7		8.1		21.2		16.8
2,000	−28.66		92.4		114.3		48.7		−20.5	
		5.09		12.3		7.6		19.9		15.2
3,000	−23.57		104.7		121.9		68.6		− 5.3	
		5.09		11.1		7.2		18.8		13.8
4,000	−18.48		115.8		129.1		87.4		+ 8.5	
		5.08		10.2		6.9		17.7		12.7
5,000	−13.40		126.0		136.0		105.1		21.2	
		5.08		9.4		6.6		16.9		11.7
6,000	− 8.32		135.4		142.6		122.0		32.9	
		5.07		8.7		6.3		16.1		11.1
7,000	− 3.25		144.1		148.9		138.1		44.0	
		5.06		8.1		6.1		15.5		10.6
8,000	+ 1.81		152.2		155.0		153.6		54.6	
		5.06		7.5		5.9		15.1		10.2
9,000	6.87		159.7		160.9		168.7		64.8	
		5.05		7.0		5.7		14.8		9.9
10,000	11.92		166.7		166.6		183.5		74.7	
		5.04		6.6		5.4		14.5		9.6
11,000	16.96		173.3		172.0		198.0		84.3	
		5.03		6.2		5.2		......		9.3
12,000	21.99		179.5		177.2		....		93.6	

* 1 atm. = 1.033 kg. per sq. cm.

As regards the magnitude of the effect of uniform pressure in raising the melting-point of a pure silicate, there are no experimental data; but the initial slope of the melting-curve could be calculated if the latent heat and volume change on melting were known. The relevant evidence is very scanty and unsatisfactory; but, so far as it goes, it indicates that the latent heat is of the order of 100 cal.[1] per gram and that the volume change ranges up to 10

[1] With regard to this estimate see: J. H. L. Vogt, *Die Silikatschmelzlösungen*, II (1904), 65; W. P. White, *Am. Jour. Sci.*, XXVIII (1909), 486 (footnote); G. Mulert, *Z. anorg. Chem.*, LXXV (1912), 238.

per cent,[1] or about 0.04 cc. per gram. On this basis[2] the rise of melting-point of a *pure* silicate produced by a pressure of 1,000 atm. would range up to 15°, with a possibility of a somewhat greater rise in exceptional cases (specifically, whenever the latent heat is small, unless the change of volume is at the same time very small). Accordingly, the change of melting-point would be of the order of 6° for the first mile in depth, an amount which, in all probability, would with each succeeding mile decrease steadily at a steadily decreasing rate, but would not become zero.

Now the question arises—could the temperature gradient downward overtake the pressure gradient? It would seem that this possibility is not to be excluded, for, according to the best evidence (likewise very scanty), the increase in temperature for one mile in depth is about 40° C.; consequently, if we assume that this gradient and the melting-curve gradient remain constant, the temperature would at a depth of 35 miles be 1,400° C., the approximate temperature at which many pure silicates exposed to the pressure existing at that depth would melt.

The foregoing reasoning applies only to a pure silicate melting to a liquid of its own composition; but to reason from the behavior of a single pure substance to its behavior in contact with a complex solution is of doubtful utility. For in the latter case the relative *solubility* of the substance in the solution under the particular conditions obtaining is the decisive factor which determines whether it shall separate, though of course it will not separate unless it would normally be solid (crystalline) at the prevailing temperature and pressure; and, so far as experience goes, pressure has very little influence on solubility in condensed systems (that is, systems without a vapor phase). The foregoing serves to emphasize the fact that any conclusions as to what will happen to rocks at high temperatures and pressures must still be considered to be subject to very serious limitations.

Reversible transformation points solid $\rightleftarrows$ solid (that is, enantiotropic inversions) are altogether analogous to melting-points, and

[1] See Day, Sosman, and Hostetter, *Am. Jour. Sci.*, XXXVII (1914), 1; cf. the results cited by R. A. Daly, *Am. Jour. Sci.*, XV (1903), 276.

[2] $\Delta T = \frac{T\Delta V}{41.3\Delta H} \times \Delta P = \frac{1,500 \times 0.04}{41.3 \times 100} \times 1,000 = 15°$ for 1,000 atm.

consequently the reasoning with regard to the shape of the melting-curve is, *mutatis mutandis*, applicable here; with this difference, however, that a maximum on the transformation curve is possible, since there appears to be no necessity that the solid form stable at the higher temperature should always have the greater specific volume. On the other hand, nothing whatever can be predicted as to the influence of pressure on monotropic (apparently irreversible) transformations, such for example as marcasite→pyrite, or aragonite→calcite; for such inversions take place whenever and wherever their rate becomes appreciable. Consequently the temperature region within which a monotropic inversion takes place varies with the conditions, especially such as the presence of a solvent; but there is no evidence that uniform pressure accelerates such transformations at all.[1] This conclusion is borne out by the fact that high pressure acting on either marcasite or aragonite was absolutely without effect in inducing the transformation to the more stable form. In this connection, however, it is to be remarked that it is possible that an inversion which is monotropic at one pressure may at another pressure be enantiotropic; no actual case of this is known, but this is not surprising in view of the small amount of work hitherto carried out under a wide range of pressures.

Very frequently rise of temperature causes one crystal form to change over into another (e.g., quartz→tridymite→cristobalite); rise of pressure (at constant temperature) in many cases produces a precisely analogous result—in other words, certain modifications are really stable only at high pressure. It follows that investigation of pure substances over a wide pressure range will frequently result in the discovery of hitherto unknown modifications of those substances. A beautiful example of this is the substance water, of which no fewer than five solid modifications are now known;[2] the relations between these are given in Table II and also in Fig. 1,

[1] The influence of pressure on rate of reaction in condensed systems has been investigated in very few cases: apparently it is very small, and may be either an acceleration or retardation. See E. Cohen and R. B. de Boer, *Z. physik. Chem.*, LXXXIV (1913), 41; E. Cohen and H. F. G. Kaiser, *Ibid.*, LXXXIX (1915), 338.

[2] P. W. Bridgman, *Proc. Am. Acad.*, XLVII (1912), 439–558; for other examples see Bridgman, *Physic. Rev.*, III (1914), 153–203.

which shows the fields of stability of the several forms and the several melting and transformation curves. From the diagram it is evident that if we keep the temperature constant at $-20°$, and increase the pressure gradually, the following sequence of events will occur: at 2,000 atm. ice I will melt, but the liquid will freeze again to ice III at 2,500 atm.; this in turn will pass over into ice V at 3,500 atm. which at 6,300 atm. will pass over into ice VI. At any temperature above 0°, no solid would appear until the pressure exceeded 6,400 atm. when ice VI would appear at a pressure depending on the temperature, e.g., 20,000 atm. at $+73°$.

TABLE II

NUMERICAL DATA PERTAINING TO THE EQUILIBRIUM DIAGRAM FOR WATER—LIQUID AND FIVE SOLID FORMS* (AFTER BRIDGMAN)

Pressure kg. per sq. cm.	Corresponding Temperature	Phases in Equilibrium
1	0.0	Freezing-point: ice I–liquid
1,000	− 8.8	" " " "
2,000	−20.15	" " " "
2,115	−22.0	Triple point (*A*): ice I–ice III—liquid
2,170	−34.7	" " (*B*): " " —ice II
3,510	−24.3	" " (*C*): ice III–ice V—ice II
3,530	−17.0	" " (*D*): " " —liquid
4,500	−10.2	Freezing-point: ice V–liquid
5,500	− 4.2	" " " "
6,380	+ 0.16	Triple point (*E*): ice V–ice VI—liquid
8,000	+12.8	Freezing-point: ice VI–liquid
12,000	+37.9	" " " "
16,000	+57.2	" " " "
20,000	+73.6	" " " "

* The notation ice IV was not used, as it had already been assigned by Tammann to another reputed form, which differs little from ordinary ice (I): the existence of this form, which was not encountered by Bridgman, is very doubtful.

Nor must it be supposed that the foregoing example is exceptional; indeed, it is now certain that polymorphism—i.e., the ability of a single chemical substance to appear in more than one crystalline form—is of frequent occurrence, even without the intervention of change of pressure. For example, sulphur exists in at least four forms, silica in at least six forms, etc. Now the form stable under high pressure may persist at ordinary pressure, just as high-temperature forms may persist at ordinary temperature; we may expect, therefore, that extended pressure investigations

will yield substances hitherto unobtained or even unknown, some of which might prove to be of technical importance. An illustration of this is the new form of phosphorus recently discovered by Bridgman;[1] this form is grayish black like graphite, has a high density (2.69, as against 2.34 for the red variety and 1.9 for the yellow), is ignited with difficulty by a match, and cannot be exploded by a blow from a hammer, as red phosphorus can. Never-

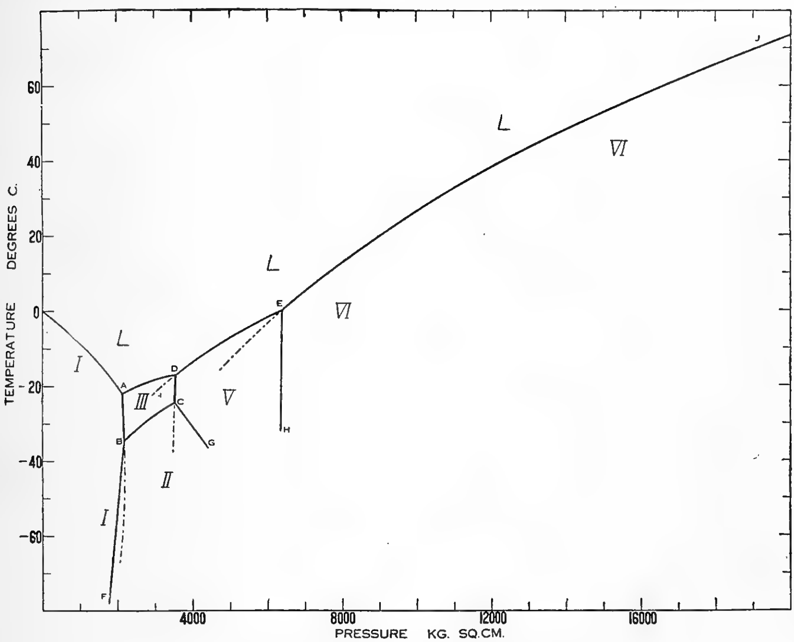

FIG. 1.—Equilibrium diagram for water—liquid and five solid forms

theless it is probable that uniform pressure has been essential to the formation of but few ordinary minerals, except in so far as it was effective in securing an adequate concentration of some volatile component (or components) in the magma from which the minerals separated.

So far we have taken up the effects of pressure upon a single pure substance; when we deal with systems of more than one

[1] P. W. Bridgman, *Physic. Rev.* III (1914), 187.

component, the mutual solubilities[1] enter as factors in the result. These are affected by pressure because it changes the relative concentrations or activities of the several components. This influence is slight with solids and liquids, because their compressibilities are very small; but in systems with a gaseous phase it is very great, because the concentration of a gas is directly proportional to the pressure, so that in such systems change of pressure will have a marked influence on the relative concentrations of the several reacting components.

In accordance with this, the influence of uniform pressure upon solubility and upon chemical equilibrium in condensed systems—that is, systems in which there is no vapor phase—is slight, and for practical purposes negligible in comparison with the influence of temperature or change of composition of the solution. Thus the few trustworthy experiments made hitherto show that a pressure of 1,000 atm. changes the solubility of a salt in water by only about 1 per cent of its value—a change which may readily follow a temperature change of a few degrees or a slight change of composition of the solution.[2] Again, at ordinary temperatures the liquids water and methylethylketone are only partially miscible at 1 atm., but they become completely miscible at about 1,100 atm.,[3] whereas at 1 atm. this does not take place until a temperature of 152° is reached. The opposite effect—i.e., unmixing by increase of pressure—could equally well be produced, just as it may be by increase of temperature. But it seems probable that neither of these effects is of importance in liquid mixtures of silicates, because liquids which

[1] It is to be remembered that mutual solubilities can be altered by the addition of a third component, even though it forms no part of the solid phase which separates. Thus, as is evident from fig. 6 of Rankin's paper (*Am. Jour. Sci.*, XXXIX [1915], 1–78), a mixture composed of 60 per cent SiO_2, 40 per cent Al_2O_3, on cooling deposits Al_2SiO_5 (sillimanite), whereas in presence of CaO, in amount ranging from 3 to 30 per cent of the whole, the solid phase which separates is *pure* Al_2O_3. This illustrates how even a small change in composition of a solution may alter the course of crystallization.

[2] E. Cohen and L. R. Sinnige, *Z. physik. Chem.*, LXVII (1909), 432; LXIX (1909), 102; E. Cohen, K. Inouye, and C. Euwen, *ibid.*, LXXV (1911), 257. These authors give a critical résumé of earlier work along this line.

[3] P. Kohnstamm and J. Timmermanns, *Proc. K. Akad. Wetenschappen*, XV (1913), 1021–37.

are so much alike chemically as are the silicates are usually completely miscible under all conditions.

But the effect is of another order of magnitude in presence of a gas phase which may enter into reaction or solution with other components; and consequently this circumstance must be taken into consideration in any discussion of the behavior of the magma, which undoubtedly contains volatile components of this character.

TABLE III

THE SOLUBILITY* OF $CaCO_3$ AT 16° AND ITS DEPENDENCE UPON THE PARTIAL PRESSURE OF CO_2 OVER THE SOLUTION

Partial Pressure of CO_2 in Atmospheres	Solubility Parts $CaCO_3$ per Million Parts Water
0.0	14.3
0.00037	46.1
0.00050	74.6
0.0033	137.2
0.0139	223.1
0.0282	296.5
0.0501	360.0
0.142	533.0
0.254	663.0
0.417	787.0
0.553	885.0
0.730	972.0
0.984	1,086.0

* Data according to Schloesing (*Compt. rend.*, LXXIV [1872], 1552; LXXV [1872], 70, except the first two, which are due to Kendall [*Phil. Mag.*, XXIII (1912), 958–76]).

As a simple example, consider the amount of $CaCO_3$ which will dissolve in water inclosed in a vessel with various partial pressures of CO_2. In pure water it is 14 parts per million, an amount which is trebled by the presence of as little as 0.00037 atm. CO_2 (3.7 parts in 10,000, about the proportion of CO_2 in ordinary air) and has increased by more than 70 times when the CO_2 pressure reaches 1 atm. (see Table III). This great increase is an example of the very great effect of increase of concentration (partial pressure) of a gaseous component; it is, of course, due to the formation of calcium bicarbonate in the solution, but this circumstance in no wise affects the force of the argument. Indeed, this is an excellent illustration

of the law of mass action, as the calculations made by Stieglitz show.[1]

A closely related point is that a compound resulting from a reaction involving a volatile component will not be formed unless the concentration—in other words, the partial pressure—of that component is above a certain limiting value, this value being dependent on the conditions, on the temperature especially. An illustration may serve to make this clearer. If we wish to keep liquid water at 200° we must have it in a closed vessel capable of withstanding the vapor pressure of water at that temperature, which amounts to about 15 atm. (225 lbs. per sq. inch); likewise at 300° the vessel would require to be strong enough to support an internal pressure of 100 atm.; but if the vessel should leak, either because it was not strong enough or because it was not tightly closed, the vapor would escape and the liquid water would disappear. Similarly we can keep $Ca(OH)_2$ as hydroxide at 550° in steam at 1 atm., but at 750° it would be necessary to have a pressure of water vapor of about 15 atm. (and at 950°, about 100 atm.) if we wish to retain it all as hydroxide and to have no oxide present; in other words, if we should wish to prepare $Ca(OH)_2$ from CaO at a temperature of 750°, we could do so only by having continuously a pressure of at least 15 atm. of water vapor above it.

It may happen, therefore, that the product which actually separates in certain cases will be determined by the magnitude of the pressure of water vapor at the time of separation. This serves to account for the fact that the biotite of many deep-seated igneous rocks is replaced in their effusive forms by olivine and leucite (which are in the aggregate chemically equivalent to biotite from which water has escaped)—a relation which has been brought out by Iddings, Bäckstrom, and others. Similarly, muscovite is found only in granitic rocks consolidated at depth, while in their surface equivalents the water has in part escaped and the potash has entered feldspar and biotite.

[1] J. Stieglitz, "The Relations of Equilibrium between the Carbon Dioxide of the Atmosphere and the Calcium Sulphate, Calcium Carbonate, and Calcium Bicarbonate of Water Solutions in Contact with It," *Carnegie Institution of Washington, Publication No. 107* (1909) (*The Tidal and Other Problems*, by T. C. Chamberlin *et al.*), pp. 235–64.

A noteworthy point in this connection is that water vapor—or, for that matter, any volatile component—may, and does, act in a way precisely analogous to any non-volatile component, the only difference being that it in general requires pressure in order to hold it on the job, so to speak. As an illustration, let us consider the equilibrium diagrams in Fig. 2, I for the system $LiNO_3$–KNO_3, II for the system H_2O–KNO_3, III for the system H_2O–CrO_3. In I, *L* is the melting-point of pure $LiNO_3$, which by admixture of KNO_3 is lowered along the line *LE:* this line therefore represents the equilibrium between solid $LiNO_3$ and liquid mixtures of $LiNO_3$ and KNO_3 of various compositions; in other words, it is the solubility curve of $LiNO_3$ in these mixtures. Similarly *KE* represents the equilibrium between solid KNO_3 and liquid. These two curves meet in *E*, the so-called eutectic point, at which point (132°) a liquid of the composition 45 per cent $LiNO_3$, 55 per cent KNO_3 freezes to an agglomerate of the same composition. In II, point *I* is the freezing-point of water, and *IC* the freezing-point of solutions of KNO_3; in other words, *IC* is the equilibrium line along which ice is in equilibrium with mixtures of H_2O and KNO_3 of various compositions. The line *KC* represents the equilibrium between solid KNO_3 and mixtures of KNO_3 and H_2O; the end nearer *C* is the ordinary solubility curve of KNO_3 in water, while toward *K* it would be more logically called the solubility curve of H_2O in KNO_3. The curves meet in *C*, which in the case of aqueous solutions is usually called the cryohydric point. These two systems are thus obviously altogether similar, the sole difference resulting from the difference in the vapor pressures of H_2O and $LiNO_3$ in relation to the prevailing atmospheric pressure; namely, that whereas at the temperatures concerned the vapor pressure of $LiNO_3$ is inappreciable, that of water ranges from some millimeters up to many atmospheres. The equilibrium curve (III, Fig. 2) for the system H_2O–CrO_3 is included because in form it resembles very closely that for $LiNO_3$–KNO_3; the branch of the curve on the water side is much longer than in the case of H_2O–KNO_3, corresponding to the much greater solubility of CrO_3 in water at and below 0°.

At each point on the curve *CK* the liquid in equilibrium with solid KNO_3 has a definite composition and therefore a definite vapor

pressure of water which must be maintained in order to secure the appropriate concentration. This vapor pressure becomes equal to 1 atm. at about 120° (the ordinary boiling-point of a saturated aqueous solution of KNO_3); increases, as we ascend the curve, to a maximum, which in this case is only 2 or 3 atm., and finally decreases to zero at the point *K*. In systems such as this—namely, in which the liquid may have (under the appropriate conditions)

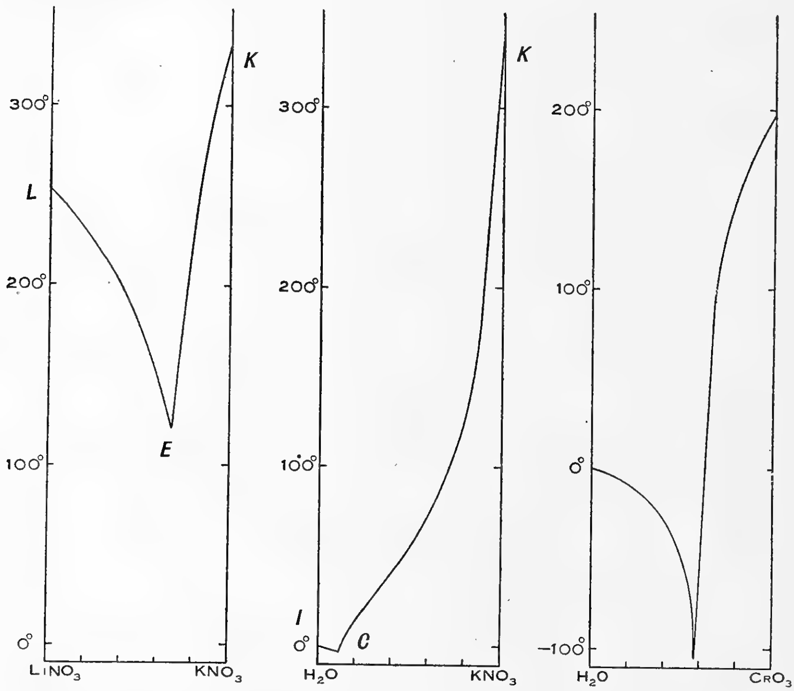

FIG. 2.—To show the similarity between the equilibrium diagrams of the system $LiNO_3$–KNO_3, H_2O–KNO_2, H_2O–CrO_3.

any composition ranging from one pure component to the other—the magnitude of the maximum equilibrium pressure depends mainly on the melting-point of the salt; where the melting-point is low, as for KOH, the maximum pressure is less than 1 atm.; where it is high, as for potassium silicate, the maximum is some hundreds of atmospheres. Nevertheless, as is evident from the diagram, the lowering of melting-point produced by a comparatively

small amount of water may be quite large; for it would seem that a lowering of 100° could easily result from the presence of water vapor at a pressure less than 20 atm. Incidentally it may be pointed out that the circumstance that pure water has a critical point at 370° and 200 atm. is of secondary importance in the discussion of systems belonging to this type.[1]

The great similarity exhibited by these two systems shows that there is no real difference between melting-curves and solubility curves; a simple and well-known illustration of this is the fact that the melting-point of $CaCl_2 6H_2O$ to a liquid of its own composition is merely a particular point on the graph representing the solubility of $CaCl_2$ in water. In general it is much simpler to consider all such phase diagrams as solubility diagrams, as by so doing it is easier to grasp their full significance and to predicate what will happen under specified circumstances. This point was clearly stated thirty years ago by Guthrie, the first investigator of eutectics, who wrote as follows:[2]

> [My experiments show] that water at a high temperature may not only play the part of a solvent in the ordinary restricted sense, but that there is in many cases no limit to its solvent faculty; in other words, that it may be miscible with certain rocks in all proportions: that solution and mixture are continuous with one another. And this continuity, as my experiments prove, is established in some cases—and these indeed with bodies having no chemical affinity with water—at temperatures not above the temperatures of fusion of those bodies *per se*.

Moreover, Guthrie saw the geological significance of these results, for he wrote:

> Just as in the selective formation of salt-alloys we may have the artificial type of the genesis of many primary rocks and metamorphic modifications, so in the wonderful solubility in or miscibility with water of such alloys and of some salts at high temperatures we may have a no less clear type of the formation of certain volcanic rocks and an explanation of some of their peculiarities. . . . . Obsidian melted and under pressure will, I presume, mix freely with water. When this pressure is gradually removed, water vapor escapes, and although it takes with it a large amount of heat, the temperature of the obsidian

[1] With regard to this point see Morey and Niggli, *Jour. Am. Chem. Soc.*, XXXV (1913), 1089 ff.

[2] *Phil. Mag.*, XVIII (1884), 117.

may still remain above its point of anhydrous fusion, or it may be maintained fused by heat from other sources. Finally cooled, it is the familiar glassy amorphous mineral. A quick release of pressure entails a quick vaporization of water and a quick loss of heat. The obsidian mass, during and because of the loss of water and the loss of heat, becomes pasty and "rises" like dough during fermentation, and becomes pumice, which is often found overlying obsidian. A quicker release of pressure from above causes the vesicular and vesiculating masses to be projected, and if the vesiculation is carried far and fast, volcanic dust is produced.

The phenomena discussed in the paragraph just quoted have now been realized experimentally with potassium silicates by Morey, who was able to produce at will either a hard or a pumiceous glass merely by altering the mode of cooling.[1]

What has been said about systems containing volatile components may be applied to elucidate some aspects of the behavior of of a cooling magma. If the crystals separating out initially from a magma situated within a confined space contain none of the volatile component, then the concentration of the volatile component in the residue would become continuously greater, and consequently the vapor pressure would *increase;* this increase might under certain circumstances be large, so that there would be a considerable tendency to enlarge the space within which the magma is confined. Accordingly eruption of a magma may be correlated with a comparatively early stage of its crystallization. On the other hand, the initiation of crystallization may be due to the circumstance that the pressure was relieved—by faulting in the adjacent rock, or otherwise—whereupon the magma began to lose its volatile components, and consequently to crystallize; this process is precisely analogous to the crystallization of a salt from a solution by boiling off the solvent.

If no escape is possible, the residue from the crystallization of the main portion of the non-volatile constituents will be a *fluid* (as distinct from a liquid) solution, containing silicates and probably sulphides, etc., which is so mobile that it can easily penetrate the adjacent rock, producing the phenomena of contact metamorphism and injection. Thus we can account for the very thin veins and dikelets often observed at igneous contacts, which indicate that at

[1] G. W. Morey, *Jour. Am. Chem. Soc.,* XXXVI (1914), 226–27.

some period the rock was permeated by very mobile solutions from which crystals subsequently separated while the still volatile components passed on. It was presumably in some such way that the pegmatites formed, for there are other reasons for believing that they formed at about 600°, a temperature at which all their constituent minerals would long have been solid had there been no volatile components present.

SUMMARY

In the foregoing pages is presented a brief discussion of some aspects of the influence of pressure on the formation of rocks and minerals. In general it would seem that the importance to geology of the effects of pressure upon so-called physical changes (e.g., the melting-point of a single pure substance) has been overestimated relatively to that of the influence of pressure upon chemical changes —in other words, upon equilibrium in polycomponent systems. Change of effective pressure will in general change the configuration of the various fields of stability in a system, acting thus in a way precisely analogous to change of temperature, or of gross composition; but in the case of pressure the effect will usually not be especially marked unless one or more of the components is volatile—that is, unless the concentration of one or more of the components really changes appreciably with change of pressure. This is merely an example of the general rule that the magnitude of the effect of pressure on a system depends upon the difference in compressibility of the several phases present, being greatest when this difference is greatest, and conversely.

Accordingly we must, in any discussion of the course of crystallization from a complex magmatic system, take into account the mode in which the effective pressure varies as well as the mode of cooling. For change of pressure, like change of temperature, may affect the order of crystallization—and even the character—of the minerals which separate; this result of course depending merely upon the circumstance that the saturation limits (solubilities) of the several solid phases which could possibly separate out are not all affected equally by change of conditions.

January 29, 1915

TWO GLACIAL STAGES IN ALASKA[1]

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The existing glaciers of Alaska are of such wide distribution and are such striking scenic features that they have attracted the attention of many glacialists. The coastal glaciers, many of which are of large size and relatively accessible, have been most frequently visited and described, and the literature on Alaskan glaciers now includes a large number of titles. The fact has been generally recognized by those who have studied the glaciation of this territory that the area at present covered by glaciers is much smaller than the area which was formerly covered by glacial ice, and glaciers no longer exist at many localities where the surface still shows conspicuous evidence of their former presence. This evidence consists of such physiographic features as typical glacial mountain valleys, U-shaped in cross-section; glacial lakes; moraines and hanging tributary valleys, and of such other evidence as is afforded by smoothed, polished, and striated rock surfaces; glaciated pebbles and bowlders; erratic bowlders, and deposits of glacial till.

It is beyond the scope of this paper to outline the areas in Alaska which are known to have been formerly covered by glacial ice. Brooks[2] published in 1906 a map showing the limits of glaciation, as known at that time. Tarr and Martin[3] have recently published a revision of Brooks's map, based on somewhat more recent information. On both of these maps a number of areas have been shown as glaciated about which no detailed information is available, and the outlines as given can be corrected only after much more field work has been done.

[1] Published by permission of the Director of the U.S. Geological Survey.

[2] A. H. Brooks, "Geography and Geology of Alaska," *U.S. Geol. Survey Paper 45*, 1906, Pl. XII.

[3] R. S. Tarr, and L. Martin, *Alaskan Glacier Studies*, National Geographic Society, 914, Map 1.

All attempts to map the limits of glaciation in Alaska have been made upon the assumption that the ice reached its greatest extension during the last great glacial advance, the evidence of which is so conspicuous, and that the limits of glaciation as shown are the limits reached by the ice during this period of expansion. Up to the present time no facts had been obtained which would show the age of this last great ice advance, as compared with any of the various stages of continental glaciation. Furthermore, although by analogy one might expect that the glaciers of Alaska would have been influenced by the same general climatic conditions which affected the main body of the continent, and would have advanced and retreated contemporaneously with the continental glaciers, yet for a period of fifteen years, during each summer of which geologists have been in the field observing glacial phenomena, the problem of whether or not there have been recurrent glacial stages in Alaska has continued to present many uncertainties. In 1890, and again in 1891, I. C. Russell[1] observed on the southern slopes of Mount St. Elias certain elevated marine deposits of fine clastic sediments containing bowlders which he believed to be of glacial origin. The same terrane was observed, in 1913, by A. G. Maddren[2] in the Yakataga district, and his interpretation of the origin of the bowlders is in agreement with that of Russell. The published literature for the most part, however, fails to discuss the probability of earlier stages of glaciation in Alaska, although the writer[3] has suggested that there may have been earlier glacial advances, but if so, they were less extensive than the last, and the traces of such glacial advances were destroyed or obscured by the more extensive and more recent ice invasion.

During a geologic reconnaissance trip into the White River basin, in the summer of 1914, observations were made which seem to throw light both upon the age of the last great ice advance and

[1] I. C. Russell, "An Expedition to Mount St. Elias," *Nat. Geog. Magazine,* III (1891), 170–73; also "Second Expedition to Mount St. Elias in 1891," *13th Ann. Rept. U.S. Geol. Survey,* Part II (1893), pp. 24–26.

[2] A. G. Maddren, "Mineral Deposits of the Yakataga District," *U.S. Survey Geol. Bull. 592,* 1914, pp. 131–32.

[3] S. R. Capps, "The Bonnifield Region, Alaska," *U.S. Geol. Survey Bull. 501,* 1912, pp. 35–36.

upon the problem of whether or not there have been recurrent glacial stages in Alaska. The first of these problems has been discussed elsewhere,[1] and the mere statement will suffice here that the last great ice advance was probably contemporaneous with the Wisconsin continental glaciation.

Near the source of White River in Russell Glacier and lying between Lime and Solo creeks, two of its tributaries, there are certain foothills of the mountains which in 1908 were seen by the writer to consist for the most part of gravels, but no careful study of this section was then made. In the summer of 1914 a

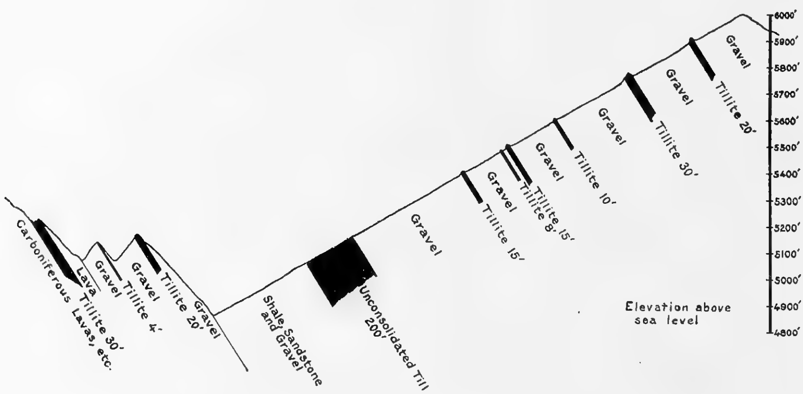

FIG. 1—Diagrammatic section of older glacial deposits near head of White River

single day was available in which to revisit this locality, and heavy rains and fog during that day prevented as thorough an examination as was desirable. The section shown in the accompanying figure (Fig. 1) is therefore not complete and the thicknesses given are only approximate, but nevertheless certain important facts are made clear. In the section to be described the exposure is unusually good, the surface of the hills being almost entirely free from vegetation, and cut by a number of clean, steep-sided gullies so that practically every foot of the deposit is exposed (Fig. 2).

The section, covering a vertical range of 1,150 feet, shows a great thickness of unconsolidated gravel beds, with some soft shales

[1] S. R. Capps, "An Estimate of the Age of the Last Great Glaciation in Alaska," *Jour. Wash. Acad. Sci.*, V, No. 4 (1915), 108–15.

FIG. 2.—The older glacial deposits form the hill near the center of the picture. The lenticular beds, which project above the general slope, are lenses of tillite, separated by unconsolidated gravels and unindurated till.

and a little sandstone, interrupted by sheets and lenses of glacial till in varying stages of induration, and by lava flows. Measured perpendicular to the dip, more than 3,000 feet of beds were examined, and the thickness of the upward continuation of the series is not known, but is considerable. The gravels, which comprise much the greater part of the entire thickness, are well rounded but only fairly well assorted, and are apparently stream-laid. The pebbles are composed of the Carboniferous lavas and limestones which form the mountains immediately adjoining to the southwest. In one place a considerable thickness of arenaceous shales and sandstones occurs. Only one lava bed is shown in Fig. 1, but in other near-by localities similar lenticular lava flows were seen much higher in the section, and farther east, on North Fork of White River, the glacial series here described is overlain by lavas of considerable thickness.

The tillite beds, while forming only a small proportion of the whole series, are nevertheless the most conspicuous members of it, for their outcrops are left in high relief by the removal of the softer surrounding materials. The uppermost bed shown in the figure in places forms the crest of the ridge and stands up in high ragged pinnacles (Fig. 3). The tillite exactly duplicates, except for its induration, the ordinary glacial till which is of such widespread distribution throughout this general region. It has a clayey matrix full of small, angular particles of rock and incloses abundant pebbles, bowlders, and angular fragments of rock, many of which are several feet in diameter. The included bowlders and blocks are of the materials which compose the mountains to the southeast and are mostly of basic extrusive rocks of brown, purple, and reddish color, and the matrix has a slight purple tinge.

Striae were found abundantly on many bowlders, especially on those of fine texture. Large bowlders in particular showed plentiful striations, but typically striated hand specimens were not easily found. The characteristic subangular bowlders, so typical of glacial deposits, are plentiful throughout the tillite, and the whole aspect of these beds leaves no doubt of their glacial origin. A number of striated and subangular pebbles were broken from the hard matrix, and numberless other larger, but equally characteristic,

striated and subangular bowlders were found imbedded in it, so that no suspicion can be entertained that the glaciated bowlders were deposited upon the surface at this place by a later glacier.

FIG. 3.—Close view of a tillite bed

In the particular part of the section measured, and shown in Fig. 1, there are nine beds ot tillite, ranging from 4 to 30 feet in thickness, and one bed of uncemented glacial till, much covered by waste from above, but apparently at least 200 feet in thickness. Probably no other portion of the section would be like that figured,

for the tillite beds are lenticular in the cross-section exposed and probably also down the dip. The uppermost tillite bed shown is, however, known to persist along the strike for at least a mile, though its thickness is variable.

On North Fork of White River, about 6 miles east of the locality here described, similar tillite beds are exposed in the canyon of that stream and are there overlain by lava flows. The lava bed shown near the base of the section and other lava flows interbedded with the gravels and tillite higher in the series, but not at this particular place, are brightly colored, reddish, scoriaceous lavas, like the flows which are so generally distributed throughout the Wrangell Mountains. They range in age from Tertiary to Recent, and Mount Wrangell even now shows signs of mild activity.

As already stated, the evidence seems fairly definite that the last great ice advance in the White River valley was contemporaneous with the Wisconsin stage of continental glaciation. Till left by the ice during this advance is of widespread distribution in the White River basin, and wherever observed is unconsolidated and little oxidized. At one point at the edge of the Lime Creek gravel flat, about 2½ miles from the locality of the section shown, the unconsolidated till of the last ice advance was found lying unconformably upon the upturned and glaciated edges of the older tillite, proving definitely that the tillite series was laid down during a glacial advance which antedated the last ice invasion. Furthermore, a long time-interval between the two glacial advances is indicated by the induration of the tillite and by its deformation since it was deposited. The physiographic evidence at hand also goes to show that the older tillite was indurated and uptilted and a mature topography eroded upon it before the last great glacier remodeled the surface and left its morainal deposits.

The tillite series here discussed is thought to have been laid down near the oscillating edge of a glacier. The record shows repeated advances, with deposits of morainal material, followed by recessions during which the till was covered by outwash gravels from the ice front, and one period during which finer sediments, represented by the shales and sandstones, were laid down. These materials may represent lake deposits. There were also occasional

extrusions of lava over the surface of till and outwash. The lenticular shape of the tillite beds may be due either to the shape of the original morainal beds, or to local erosion following a time of ice recession. The lack of induration of one thick till bed may have been due to the impenetrability of the till itself, or of the underlying shale, and is not believed to affect the general interpretation of the section. The whole series of beds, with the exception of the lava flows, is probably similar to the deposits now being laid down near the terminus of many a glacier in this same mountain range.

Little can now be said of the areal extent of the glacier which left these old morainal deposits. The elevation of the present exposures is of little importance, for the beds show by their structure that they have been tilted, with minor folding, and their present elevation may be much different from that at which they were laid down. At present the 3,000 feet of the beds described are seen within a vertical range of only 1,150 feet. The beds at this locality dip from 55° to 60° to the east, but the dips gradually become less as the distance from the mountains increases, and as seen in the canyon of North Fork of White River the tillite is nearly flat-lying. All the known outcrops of tillite lie well within the limits reached by the ice during its last great advance, and no comparison can yet be made of the extent of the ice fields during the two stages.

Summary.—There have now, for the first time, been found in Alaska deposits of glacial till which can be proved to antedate by a considerable period of time the last great ice expansion, thus proving that there have been at least two distinct glacial stages in that territory. The deposits comprise a series having a thickness over 3,000 feet and consist of indurated as well as unconsolidated glacial till sheets, separated by outwash gravels and some assorted sediments, and interrupted by lava flows. The section examined was evidently deposited near the border of an oscillating ice edge and shows ten definite advances of unknown magnitude, represented by the deposition of till beds, followed by periods of retreat during which water-laid beds were deposited upon the successive till beds. After its deposition this series of glacial and glacio-fluvial beds was covered by lava flows, at least locally, was uplifted, in part indurated, and was later deeply cut by erosion. A much later ice

advance, probably contemporaneous with the Wisconsin continental glaciation, then took place, overriding the earlier glacial deposits and locally capping them with later till beds. The extent of this earlier glacial advance, as compared with the last great stage of glaciation, is not known, all of the observed deposits of older glacial material lying far within the outer limit of the last great glaciation. Enough data have not yet been obtained upon which to base a correlation of the older glacial deposits with the earlier stages of continental glaciation.

STRATIGRAPHY OF THE WAVERLY FORMATIONS OF CENTRAL AND SOUTHERN OHIO (*Continued*)

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PART II

CUYAHOGA FORMATION (*continued*)

THE SCIOTO VALLEY SHALE FACIES

Extent and Thickness.—The Scioto River from Chillicothe to Portsmouth flows roughly along the axis of this area. Here the Cuyahoga is constituted almost entirely of bluish-gray clay shale with only an occasional sandstone bed. To the eastward, as the margin of the Hocking Valley conglomerate facies is approached, sandstones become more numerous and increase in thickness. A similar increase in sandstone content is experienced to the westward, as the Vanceburg sandstone facies is approached. These transitional sandstones are, however, of a different type on opposite sides of the shale area. On the eastern side are found hard, coarse sandstones (but finer grained than the Hocking Valley sandstones) that are peculiarly harsh to the touch; they are reddish-brown on the outcrop, light greenish-blue under cover, and *Spirophyton* is not present to any extent. Beds of this type persist as far west as the Scioto River, where they may be four or five feet thick, and some of the beds can be traced for several miles. On the western side of this shale facies, the sandstones are of the type which characterizes the Vanceburg facies, to be described presently, but distinguished particularly by their soft, clayey nature, light yellowish color on the outcrop but blue-gray under cover, and abundance of *Spirophyton*. One member only of the latter type persists any considerable distance into the shale area, the Buena Vista member, and it may be found in some sections east of the Scioto River in the presence of sandstones that have obviously been derived from the eastward, but even in the same section

each exhibits the characteristics of its respective type little changed.

The Cuyahoga over the shale area is not as thick as in the Hocking Valley facies. Between Chillicothe and Waverly it is about 300 feet thick and at Portsmouth about 300 or 320 feet. West of the Scioto River near Portsmouth it falls to about 260 feet.

Members.—Three members are distinguished. At the base is the *Henley shale member* which apparently attains a thickness of over 200 feet. This is succeeded by the *Buena Vista sandstone member.* Since both are recognized in the Vanceburg sandstone facies, their description is better reserved for that place.

The name *Portsmouth* is applied to the upper shale member from Portsmouth, where it is excellently exhibited on both sides of the Scioto River, and on the Ohio River above the town. At the "Two-Mile Hill" at Portsmouth it is 253 feet thick, but this decreases to the northward and westward. The Portsmouth shale is the equivalent of the three upper members of the Vanceburg facies. It is impossible, at present, to present correlations between the members of the Scioto Valley shale facies and those of the Hocking Valley conglomerate facies to the eastward. All that can be suggested is that the Buena Vista sandstone is not the equivalent of any one of the four members distinguished there; its horizon is probably near the middle of the deposits in the Hocking Valley. Further, it appears that the upper part of the Portsmouth shales at Portsmouth—how much is not known, but perhaps 50 feet or more—is the stratigraphic equivalent of the lower part of the Byer member of the Logan formation only six miles up the Ohio River at Sciotoville.

THE VANCEBURG SANDSTONE FACIES

Extent and general nature.—This is confined to the western parts of Scioto and Pike and the southwest corner of Ross counties, with some extension into Adams County. The facies proper lies well west of the Scioto River. The thickness of the Cuyahoga appears to be about 250 feet, of which the lower 150 feet or more are made up largely of sandstones, the upper 100 feet largely of shale.

The deposits which distinguish this facies differ greatly from those in the two sandstone and conglomerate facies just described. They consist of alternating beds of exceedingly regular sandstone and shale. The sandstones are usually from 6 or 8 inches to 18 inches in thickness, although they exceptionally attain 4 or 5 feet; they are very uniformly fine grained, blue-gray in color when unweathered, and persistently uniform in thickness for considerable distances. The shales not uncommonly are mere partings between these beds, but they may equal or exceed the sandstone beds in thickness. At McDermott the thin sandstone and shale beds are known to hold their characteristics individually over almost a square mile. The shales are thicker at the bottom and at the top and the persistent occurrence of a considerable amount of red shale in the lower part is one of the characteristic features of this facies and an unusual one in the Waverly.

The upper surface of each stratum of sandstone is covered by *Spirophyton* markings, the tubes of which penetrate downward through the upper 4–8 inches in infinite numbers. The under surface of each sandstone is usually traversed by many raised lines and elongate rootlike ridges in various degrees of prominence which are merely the casts made by the sand settling in the scratched and gouged upper surface of the underlying shale partings. These structures indicate a peculiar cycle of events that was repeated many times during the accumulation of the deposit: (1) accumulation of sands to a thickness of a foot or two; (2) period of quiet and invasion of the sandy floor by myriads of fixed worms (if *Spirophyton* is correctly interpreted as a worm tube); (3) deposition of a thin bed of clay mud and extermination of the worms; (4) intervention of some agent, probably bodies floating or trailing in shallow water, which scratched the muds but never, so far as observed, to a depth of over an inch; this completed the cycle and sands were again laid down. This was repeated many times but it is more striking in the lower part of the sandstones, where the shales are reduced to mere partings.

The deposit is marine, as is shown by the occasional finding of a sandstone bed which differs from the others only in the presence of a normal marine fauna of brachiopods, gastropods, cephalopods,

lamellibranchs, etc., in very great abundance often only for a few square feet; the composition and structure of such a deposit suggest that this accumulation where found was accomplished through current-action.

The deposits of this facies pass into the shales of the Scioto Valley shale facies to the eastward by disappearance of the sandstones. This transition takes place rather abruptly and practically all of the beds disappear simultaneously except those of the Buena Vista member, which persists far into the shale area. The original western margin of the Vanceburg facies in Ohio has been removed by erosion.

Five members can be recognized, of which all are persistent over the area of the facies, except the topmost, which has been removed by erosion at points in Pike and Ross counties. Although these members appear to be distinct stratigraphic units over much of the area, near the margins of the facies where the sandstones are disappearing certain of the boundaries obviously become of uncertain stratigraphic significance.

The Henley shale member is named from Henley, Scioto County. In the area of maximum development of the Vanceburg type of sandstones, it consists of alternating gray and red shales, but the red color disappears to the eastward, together with the sandstones of the facies. Along the western boundary line of Scioto and Pike counties this member gradually increases in thickness from $5\frac{1}{2}$ feet at Buena Vista and 9 feet at Vanceburg to $34\frac{1}{2}$ feet at Bainbridge in Ross County. Eastward and northeastward from this line of sections in the Scioto Valley shale facies the thickness increases rapidly, especially to the eastward. Along the Scioto River the thickness increases from $55\frac{1}{2}$ feet at Rushtown in Scioto County to 89 feet near Summit in Ross County. Several miles east of Piketon in Pike County, its thickness has increased to 226 feet, if the superjacent Buena Vista sandstone is correctly identified. The rapid westward thinning of this basal portion of the Cuyahoga is believed to indicate that the Cincinnati dome was an appreciable topographic feature on the shallow sea-floor of that time. There is, however, no physical evidence of erosion of the underlying thin Sunbury shale prior to its accumulation.

The name *Buena Vista* has been adopted for the second member in the Vanceburg facies.[1] For many years the village of Buena Vista was the center of extensive quarrying. The basis of the industry was a relatively thin sandstone member a few feet above the base of the Cuyahoga. Since the product was largely marketed in Cincinnati, the member came to be known as the "City ledge." It is to this member and its continuation over Scioto, Adams, Pike, and Ross counties that the name Buena Vista is here restricted.

The Buena Vista consists of sandstones of the type just described as characteristic of the facies. It is not, however, of the same quality, structure, and thickness in all parts of its area. Along the western margin of Scioto County it is usually from 3 to 6 feet thick and consists of from one to four distinct sandstone beds. Eastward from this and over the ever-thickening Henley shale,

[1] This name was first proposed by Edward Orton in his "Report on Pike County" (*Geol. Surv. Ohio*, II, Part I [1874], 626), where the lower 50 feet of the Cuyahoga formation was separated as the Buena Vista member. "As the most valuable of the building rock, however, that is furnished by this part of the series in southern Ohio occurs within 50 feet of the slate [Sunbury], these 50 feet next above the slate may be somewhat arbitrarily taken as a subdivision." Prosser has fully recounted the history of this usage and variation in usage of the term by Orton and has himself adopted it in the same sense as originally used (*Am. Geol.*, XXXIV [1904], 341, 342). It can readily be shown, however, that any member arbitrarily established on the lower 50 feet of the Cuyahoga over central and southern Ohio is wholly without foundation in fact. Stratigraphically there is no such member. The lowermost 50 feet will not include all of the valuable building stones, if this were an acceptable criterion for the establishment of such a member, since it does not include the stone at the only locality where such stone is now being worked, McDermott. The sediments forming the lower 50 feet from point to point are not even always similar; in some places it is sandstone, in others it is shale. It becomes necessary to redefine the term Buena Vista if it is to be used at all, and in the present statement it is applied to the "City ledge" of the quarrymen at Buena Vista, and its continuation throughout western Scioto and Pike counties. Orton has on one occasion (*Geol. Surv. Ohio*, V [1884], 602) used the term Buena Vista in this sense. Elsewhere he has used it apparently to denote the particular bed which he erroneously supposed to be the equivalent of the City ledge ("Report on Pike County," *Geol. Surv. Ohio*, II, Part I [1874], diagrams between pp. 618 and 619). Prosser, too, in an earlier publication used the term similarly (*Jour. Geol.*, X [1902], 289–91). The term thus redefined is not, then, used with a wholly new meaning.

The adoption of the term in Ohio with either of the meanings outlined antedates its application by H. D. Campbell to a formation in Virginia (*Am. Jour. Sci.*, 4th ser., XX [1905], 445–46), as pointed out by Prosser (*Am. Jour. Sci.*, 4th ser., XXI [1906], 181–82).

it increases in thickness to about 25 or 30 feet along the west side of the Scioto River well within the Scioto shale facies. Simultaneously the number of component beds increases. Farther east than this on the east side of the Scioto River, and northward over western Pike and southwestern Ross counties, its beds become more shaly and worthless. It remains, however, so far as traced, much the same type of deposit as characterizes the Vanceburg facies, even where, on the eastern margin of the Scioto shale facies, it becomes intertongued between those sandstones of a wholly different type that are clearly the western fringe bordering the Hocking Valley conglomerate facies. At the most easterly point where it was observed, however, several miles east of Piketon, the component beds are much thicker than a mile or two to the westward, and there is evidence that, as the Hocking Valley area is approached, the member increases in importance.

Westward from the western margin of Scioto County the Buena Vista member continues to thin, and within two or three miles of the zone where it is from 3 to 6 feet thick and economically important, it is less than a foot in thickness. It has not been traced westward beyond this but it probably thins out entirely. This is near the western limit of its outcrop belt.

The Rarden shale member is named from Rarden, Scioto County. It is not recognized outside the Vanceburg facies, since opportunity to identify its upper limit is lost with the disappearance of the overlying sandstones. Like the Henley member, it consists of alternating red and gray clay shales. At Vanceburg, Kentucky, it is $8\frac{1}{2}$ feet thick; from here northward and eastward it increases in thickness to 28 feet at Buena Vista, 27 feet at Rarden, 50 feet at Bainbridge, and about 58 feet at Elm Grove in Pike County on the extreme eastern margin of the facies.

Along the western margin of the outrcop belt, in Adams County, where the Buena Vista member becomes insignificant, this member probably rests on, and is indistinguishable from, the Henley member, although no section of precisely this type has been seen. At Mineral Springs, Adams County, there are something over 40 feet of red and gray clay shales at the base of the Cuyahoga, with a single 3-inch sandstone in the lower part, believed to be

the Buena Vista. This is the nearest approach to this condition observed.

The Vanceburg member consists of typical sandstones of the facies. It is best developed in the vicinity of Vanceburg, Kentucky (from where the name is derived), and Buena Vista. Here the member is about 150 feet thick and on the whole the sandstones are thicker than the intervening shales. Both northward and eastward the shales become relatively more important, and at Bainbridge they form much the larger part of the member. These sandstones give place by transition to the shales of the Scioto Valley facies, but rather abruptly, along the line indicated on the map.

The Churn Creek member is seldom seen and then not well. It consists of argillaceous shale with an occasional thin sandstone and the deposits of the Vanceburg member pass into it gradually. Except for the difference in relative amounts of sandstone and shale it is of the same type. It appears to be between 50 and 100 feet thick, but no section has been seen which admits of determination of this point. It is present in Ohio only in western Scioto and eastern Adams counties, being removed farther north, and only in the vicinity of Buena Vista is the Logan formation found overlying it. It is named from Churn Creek in the southeastern part of Adams County, on the divide between which and the head of Lower Twin Creek good outcrops are seen in roadside ditches.

THE CUYAHOGA FORMATION IN KENTUCKY

It has been shown how the outcrop belt of the Cuyahoga passes southwestward diagonally across the five successive facies from the Toboso facies in central Ohio to the Vanceburg facies at the Ohio River. From the Ohio River the outcrop belt of this formation continues southwestward into Kentucky. The writer has seen it at various points as far as the center of the state[1] but is not prepared to discuss it from such observation. Morse and Foerste have published sections from this region that indicate something of the

[1] Under grant from the Esther Herrman Research Fund of the New York Academy of Sciences.

nature of the formation,[1] but their sections are for the most part of the Bedford, Berea, and Sunbury formations. At a few localities only is more than the very base of the Cuyahoga described and in no one section is the entire formation distinguished.

From these observations it appears that there lies southwest of the Vanceburg sandstone facies in Kentucky an area in which the Cuyahoga is largely a bluish argillaceous shale, and that the margin between them lies somewhere between 5 and 20 miles south-southwest of Vanceburg, Kentucky. Not only do the sandstones that characterize the Vanceburg facies largely disappear, but the red shales that are equally characteristic are replaced by bluish ones. A sandstone member of precisely the same character as the Buena Vista member on the eastern side of the Vanceburg facies and occupying about the same position in the formation is continued to the southwestward into this shale facies. It is quarried at Farmer, Freestone, and Rockville, about 30 miles southwest of Vanceburg, where it is about 25 feet thick and is reported underlain by from 1 to 11½ feet of soft shales. Morse and Foerste consider this sandstone the top of the Buena Vista member, using that term with the sense given it by Prosser. Although it closely agrees with the Buena Vista member (sense adopted by the present writer) in structure and position, this correlation cannot, as yet, be considered established. Within 10 miles farther to the south-southwest, this member appears to pass into shales and the Cuyahoga is almost entirely a shale, but apparently with diminished thickness.

The question may well be raised whether these shales are the equivalent of the Cuyahoga formation, as it has herein been described, in southern Ohio. A careful tracing of the sandstone members in the lower part is necessary to an opinion as to whether the base is approximately the same. As to the upper limit, *Allerisma winchelli* was found by the writer in the overlying yellow sandstones, perhaps 100 feet above the shales, near Morehead, Kentucky. This species is, as far as known, confined to the Byer and Allensville members of the Logan. This would indicate that

[1] "The Waverly Formations of East Central Kentucky and Their Economic Values," *Kentucky Geol. Surv., Bull. 16* (1912), 76 pp.

the lower 100 feet at least of these sandstones in northern Kentucky, within 45 miles of Portsmouth, are the approximate equivalent of the Byer member in southern Ohio, which there overlies the Cuyahoga or in part passes into the topmost beds of the Cuyahoga.

PALEOGEOGRAPHY OF THE CUYAHOGA

The Toboso and Hocking Valley conglomerate areas.—Herrick thought that the material composing the Black Hand conglomerate came from the northeast.[1] Stevenson writes of the pebble material of the "Logan" (Orton's usage, Black Hand and Logan of this paper) as follows: "They can hardly have come from the north, for there the upper Logan, Reid's Olive Shales, is very fine in grain, while farther south it becomes coarse, as it is in northwest Pennsylvania, east from Reid's localities. It is equally improbable that the pebbles come from the east, for the deposits become finer eastward toward the central line of the basin, beyond which they become coarser. The sands must have come from the western side."[2]

The immediate source of the conglomeratic material in the Cuyahoga of central Ohio undoubtedly was to the south-southeastward. Lamb's observations in northern Ohio on conglomerates of undetermined horizon indicate the same direction as the source of the material.[3] It is not possible at present to state under precisely what conditions the conglomerate sandstones of the Toboso and Hocking Valley facies were deposited, but it is the writer's opinion that they are the result of deltal action under marine conditions. The structures observed, however, do not in any way correspond to the structures suggested by Barrell as criteria of such an origin.[4]

These sandstones were deposited under conditions of strong current-action which swept pebbles up to an inch in diameter to their present position; which laid down thick beds over wide areas at an inclination of 5° to 10° (much steeper than the foreset beds of the large deltas of the present, according to Barrell); which generated in these beds cross-bedding of the most pronounced kind,

[1] C. L. Herrick, *Bull. Sci. Lab. Dennison Univ.*, II, Part I (1887), 9, 10.

[2] J. J. Stevenson, *Bull. Geol. Soc. Am.*, XIV (1903), 91.

[3] G. F. Lamb, *Ohio Naturalist*, XIV (1914), 344–46.

[4] Joseph Barrell, "Criteria for the Recognition of Ancient Deltal Deposits," *Bull. Geol. Soc. Am.*, XXIII (1912), 377–446.

of both the irregular cut-and-fill and the regular inclined-bedding types; which, over extensive portions of the eastern side of the Hocking Valley area, dumped coarse conglomerate to a thickness up to 50 feet or more at an angle of from 15° to 20°, with northeasterly dips.

The material on the whole is far coarser than that observed, according to Barrell, at present in the deltas of large rivers, and the current-action was far stronger. It is the writer's opinion that the sands of the more gently inclined beds (5° to 10°), where cross-bedded as in the Black Hand member of the Toboso facies, were spread out on the ocean floor in part by the assistance of wave-action and shore currents, but chiefly under the direct influence of strong currents from a river mouth or other source of sand-laden waters. Where cross-bedding is reduced in amount and magnitude, as in the beds of the Fairfield member, the stream current is believed to have been relatively less important and the wave and shore currents more important. Much of the Black Hand member of the Hocking Valley facies has been laid down, according to this interpretation, under the former conditions, but large portions, especially near the margins of the facies, under the latter conditions. Where steeply inclined conglomerates with dips of 15° to 20° are extensively developed, as in the eastern portion of the Black Hand of the Hocking Valley area, it appears that the material was dumped into a protected area of relatively deep water, where it formed a face at the angle of repose for such material in water, uninfluenced by waves or bottom currents. A glance at the map will show that the eastern margin from Lancaster to Logan must have been, under the interpretation given these facies, well protected, and it is here only that such extensive steeply inclined bedding is found. It seems necessary to believe that the wave-action was far weaker than that obtaining along the seaward faces of present-day deltas facing the open ocean; otherwise this material would have been spread laterally along shore and outward to such an extent that the Toboso and Hocking Valley facies would not have remained as sharply defined, individual lobes.

The structures observed in the cross-bedded conglomerate, whether steeply or gently inclined, are precisely the ones that

Barrell regards as indicative of river-laid sands,[1] and not indicative of wave-laid or wave-current-laid sediments; but that they were accumulated under subaquatic and probably submarine conditions is indicated (1) by the rapid lateral passage of the members into marine beds; (2) by the presence of oscillation wave ripples on an inclined bedding-plane between sharply cross-bedded conglomeratic sandstones, in one case known over several acres and through a vertical range of some 30 or 40 feet; (3) by the presence of worm trails (*Scalarituba*) that are common in undoubtedly marine Waverly sediments; and (4) by the occurrence of thin beds of soft clay shale interbedded with the inclined sandstones and similarly inclined. No mud cracks have been seen and there is none of that evidence of oxidation seen in highly colored red beds, which may be the result of, and commonly are held to suggest, subaerial accumulation. To be sure, red sandstones occasionally appear, but they are not confined to one bed so much as to a limited locality where a considerable thickness of beds may be involved, and where, it appears, some other local condition or source must be involved.

The structures observed are very like those that would be expected in bars or spits built into deeper water, such as Sandy Hook, but perhaps with more cross-bedding; however, from the geographic distribution of the conglomerate masses along the belt of outcrop it is at present impossible to entertain this view. It does not appear possible for two such bars as the Hocking Valley and Toboso facies would imply to be built alongside of and parallel to each other simultaneously; and the results obtained so far all indicate that they were accumulated simultaneously.

As has been pointed out, there is distinct lobation in the Hocking Valley area; east of the Hocking River in southern Fairfield County, and traversed by it in Hocking County, is a secondary mass of conglomeratic sandstone that is immediately adjacent to the main mass of the facies, and it appears very probable that under cover of the Coal Measures to the southeastward the Toboso and Hocking Valley masses may become contiguous in the same way and prove to be only larger lobes of a yet greater development of coarse material. If the interpretation is correct that these

[1] *Op. cit.*, pp. 430–32.

sediments are of deltal origin, then the outcrop belt in central Ohio appears to be that portion of the floor of a marine interior sea which lay just off the front of a delta and was invaded by lobes from it. It appears, however, that no portion of the outcrop sediments have been deposited subaerially. The shoreline perhaps lay at no great distance to the southeastward, possibly only a few miles. The sea must have been one of quiet waters compared with the open ocean, otherwise the waves would have destroyed the fairly sharp lines of the facial boundaries.

It is, perhaps, unexpected to learn that these conglomeratic sandstones had their origin in the Appalachians. The paragraph quoted above from Stevenson's carefully considered work indicates how far it was from the thought of one who knew the whole problem. Certainly the deep-well records so far published from southeastern Ohio lead one to believe that the corresponding portion of the column under cover is largely shale. Well records, however, are very unreliable except when samples are taken, and usually but little attention is really paid by the driller to any except the few horizons that are well known to him and are used by him as guides.

It appears probable that, in certain places at least, the Cuyahoga sediments under cover in southeastern Ohio may be coarser than they have been described. However, it does not appear necessary to assume that the Cuyahoga under cover to the southeastward must be made up of similarly coarse sediments over wide areas. A great deal of coarse material, such as is found in either of these facies, might be transported through a relatively much smaller channel in fine-grained deltal sediments of an earlier stage of accumulation. To account for the conditions in this way would probably necessitate the assumption that there was a change in the nature of the material transported toward increasing coarseness, aggradation in proportion, and a consequent building out over the subaerial portion of the delta of a sandstone member of some thickness, but this thickness, it appears, would be thin in the lower courses of the delta.

The Vanceburg sandstone facies.—The Vanceburg facies was accumulated under conditions different from those of the Toboso and Hocking Valley conglomerate facies. Nevertheless, it probably

represents a phase of the same deltal action. Without much doubt the source was to the south-southeastward; it is not, however, inclined structure which indicates this, but the increase in the relative amount of sandstone in that direction. At one time the writer believed that the deposits of this facies were the offshore accumulations on the east side of the Cincinnati dome and that they were evidence that the dome at that time was land area. The rapid westward thinning of the Henley shale member at the base of the Cuyahoga supported this, and indeed, so far as the writer can see, can be interpreted only as indicating that the dome was at least an appreciable topographic feature on the ocean floor at the beginning of Cuyahoga time. It does not now appear, however, that the sandstone of the Vanceburg facies can be interpreted as evidence of a shore near by to the westward. The fact that in Kentucky, within a few miles of the Ohio River, a pure shale facies is found lying to the westward of this sandstone area with no red shale demonstrates that its general nature is that of a lobe of sandstone, pushed out into open water from the mouth of a stream or other definite source of sand, not an alongshore sand deposit. Indeed, nothing has been observed anywhere in Ohio or Kentucky which suggests that land lay to the westward in the region of the Cincinnati dome.

It is possible that local conditions, for example stronger wave- and ocean-current action, will explain the structures found in these sandstones, so widely different from those of the conglomerate facies. But better still is the assumption that the stream currents (or whatever other local current may have been the source) that fed these sands were far weaker than in the Hocking Valley and Toboso areas; the wave-action being the same, under such conditions the feeble waves of this sea might well have been better able to distribute the fine sands brought in at this point, whereas in the conglomerate areas the stronger feeding currents, with their loads of coarse pebbly sand, locally dominated over the wave-action.

The source of the material.—It is of interest to note the nature of the nearest sediments of this age which are exposed to the southeastward and southward as a possible source of the material. The Waverly formations occupy the same interval in the geological

column as the Pocono sandstone formation of the Appalachians and are generally regarded as its approximate equivalent. The Pocono sandstone is exposed in various belts in the Appalachians from the anthracite fields southwestward through central Pennsylvania, the western part of Maryland, and on both sides of the boundary line between Virginia and West Virginia to a point in Virginia south of the southern part of the last-named state.

The following paragraph is presented as a brief summary of the published accounts of the Pocono in this region and of a new interpretation of the members forming it; it appears that the irregularity in thickness and composition of the Pocono at various points in northeastern West Virginia can be understood on the theory that there was there erosion of its beds to the extent of 1,100 or 1,200 feet prior to the laying down of the Greenbrier limestone.

It appears that the Pocono in northern Virginia, western Maryland, and south-central Pennsylvania consists of four members (a fifth higher member has been reported from a single locality), of which the second one in ascending order is a prominent gray or white conglomeratic sandstone, the Purslane sandstone.[1] It appears that this sandstone-conglomerate member is the one which comprises almost the whole of the so-called Pocono in its thinned more westerly outcrops along the line between Virginia and West Virginia and in West Virginia, and that there the underlying member has heretofore generally been included with the Catskill.[2] In western and southwestern Virginia, three members have long been recognized[3] which, it appears, correspond very closely to the uppermost three in northern Virginia (excluding the seldom-observed highest member), and the massive conglomerate sandstone at the base in southwestern and western Virginia corresponds exactly with the second member or Purslane sandstone to the northward. It further appears that there is beneath this sandstone in western and southwestern Virginia a member, closely corresponding to the lowest member or Rockwell to the northward, which has been

[1] G. W. Stose, Pawpaw-Hancock Folio, No. 179, *U.S.G.S.*, 1912.

[2] Piedmont, Accident-Grantsville, Franklin, Staunton, Monterey, Buchannon Folios, *U.S.G.S.*

[3] W. M. Fontaine, *Am. Jour. Sci.*, 3d ser., XIII (1877), 115–16; J. J. Stevenson, *Bull. Geol. Soc. Am.*, XIV (1903), 29, and citations there given.

doubtfully referred both to the Pocono and to the Catskill, more latterly to the Catskill. So far as the writer is aware, this correlation between the members in northern Virginia and those seen to the southwestward has not before been suggested. If it is a correct one, then the Purslane sandstone member, the second member in northern Virginia, becomes important, not only as a very persistent member along the Appalachian belt, but as the only member of the Pocono along the line of the Appalachian mountains that is comparable with the coarse sandstones and sandstone conglomerates of the Cuyahoga along the Waverly outcrop belt in Ohio. When it is further recalled that the structure of these Cuyahoga sandstones and conglomerates proves that they were derived from the south-southeastward, and that the axes of the facies point in the same direction, the possible correlation with the Purslane becomes doubly suggestive. It remains to question whether the whole of the Cuyahoga sandstones should be correlated with the Purslane, or only the upper or Black Hand member, which alone is notably conglomeratic. The question is at present too uncertain a one to discuss, but if it be found that the Black Hand alone should be so correlated, the broad extent of the member would be the only cogent argument yet put forward for the separation of the Black Hand from the remainder of the Cuyahoga sediments.

LOGAN FORMATION

The simplicity of lithological structure and stratigraphy of the Logan formation is in marked contrast with the extreme variability which is found in the several facies of the Cuyahoga. With the opening of Logan time, the localization of conditions which resulted in widely different sediments in closely adjacent areas during Cuyahoga time gave way to quite uniform conditions over the whole of the southern Ohio Waverly belt. Extreme areal variation with no very important variation in vertical succession was succeeded by widespread areal uniformity and a threefold vertical subdivision which can be recognized from the Ohio River to the center of the state and probably well into northern Ohio.

Subdivisions and extent.—The three subdivisions of the Logan formation in descending order are as follows: Vinton member, Allensville member, Byer member.

It is possible that a fourth member should be recognized. E. B. Andrews in 1878 proposed that certain beds at the top of the Logan formation at Rushville be designated the Rushville[1] group. There are there shown $22\frac{1}{2}$ feet of rather soft argillaceous shales, gray or red in color, of a type quite different from the bulk of the sandy sediments that compose the Logan, and a pink crinoidal sandy limestone one foot thick. These beds are stratigraphically much higher than those which usually form the top of the Logan. This is because the pre-Pennsylvanian erosion surface here stands, relative to the beds it cuts, much higher than in most places. In addition, these beds were stated by Whitfield to show certain faunal peculiarities that, if verified, will distinguish them from the remainder of the Logan. The fauna has been extensively collected but no statement can be made as yet.

Although the three members listed above are found throughout the length of the Logan belt from central Ohio to the Ohio River, they are not present throughout its entire breadth, as is shown by the map (see Part I) of the known distribution of the Allensville member. In Scioto County the western limit of this bed is quite definitely known. In Pike and Ross counties its absence from the western half of the Logan belt is not demonstrated because of covered slopes and absence of outcrops, but such absence is very strongly suspected. It is in part on the presence of this member that the subdivision depends. The upper member is occasionally absent because of erosion prior to the deposition of the Pennsylvanian or Coal Measures; yet more rarely all three have been removed.

The lower and upper members are fine-grained sandstone and shale deposits, the result of quiet, uniform conditions; the Allensville bed is a very coarse sandstone, or perhaps it would better be called a fine pebble bed, the result of an interval of contraction of the basin. The whole is marine throughout, although faunal remains may be rare.

The Byer member.—This is usually a fine-grained sandstone. Its beds are seldom over 18 inches thick and may tend to split up into thinner beds. In southern Ohio, and especially in the more

[1] *Am. Jour. Sci.*, 3d ser., XVIII (1879), 137.

westerly outcrops, portions of it are better described as shaly sandstone or sandy shale. Even where it is wholly a sandstone there is enough clay present to make the sandstones soft. The color as seen on the outcrop is almost invariably yellow or buff, but under sufficient cover it is probably always a shade of blue or gray. At two points only within the area herein considered is it sufficiently massive to be quarried, at Newark and at Granville in Licking County. At the former locality the beds may attain a thickness of three or four feet.

The member is named from the town of Byer in the northern part of Jackson County where it is well shown in the railroad cuts east of the village. It agrees very closely in its upper and lower limits with the Division II or Kinderhook of Herrick, and with that portion of Prosser's Black Hand at Newark lying between Conglomerate I (the Berne member) and Conglomerate II, the latter one of the beds of the Allensville member.

The coarseness of the material is influenced appreciably by the presence below it of the conglomerate masses of the Cuyahoga, especially in Licking County. The thickness has been considerably influenced. Over the northern counties, Licking, Fairfield, Hocking, and Vinton, where conglomerate sandstones constitute the most of the Cuyahoga and the thickness is correspondingly great, the Byer member is much thinner than farther south in Ross, Pike, Jackson, and Scioto counties. Where the two conglomerate areas are thickest it is 25–30 feet thick; over the intervening Granville province, where the thickness of the Cuyahoga is somewhat reduced, it is commonly about 50 feet and at one point it reaches 80 feet. Westward and southwestward from the great Hocking Valley conglomerate area, as the thickness of the Cuyahoga decreases with its passage into shale the thickness of the Byer member steadily increases from 23 feet to at least 154 feet in eastern Pike County.

The tendency of the Byer sandstones to pass into shales toward the westward, although noticeable in Ross, Pike, and Scioto counties, is most pronounced along the Ohio River. There, between Sciotoville and Portsmouth, the lower portion, to a thickness of at least 75 feet and perhaps more, passes into shales which are not distinguishable from the body of the Cuyahoga shales of

the Scioto shale facies and which have been considered the upper part of the Portsmouth member of the Cuyahoga. This tendency on the part of the Byer sediments suggests strongly that its material was also derived from the eastward and southward, in spite of the thinning of the formation to the eastward.

It is not possible to say to what extent the member continues to thicken westward because of the disappearance of the Allensville member in that direction just where the Byer passes into shales and sandstones which are lithologically so similar to those of the Vinton member as to be indistinguishable from them.

The Byer member was formed during a period of quiet sedimentation. The shoreline lay far enough to the eastward so that the outcrop belt of central and southern Ohio was wholly outside the zone of continual shifting of sands by currents and waves, and beyond the reach of delta or sand-spit accumulations such as are seen in the Cuyahoga. Sharp contacts between beds are seldom observed where shaly sandstones make up a considerable part of the succession, the change from sandstone beds through shaly sandstones to shale beds being almost always a gradational one. Well-defined bedding-planes are more likely to be found in those portions made up almost entirely of sandstones and where the material is slightly coarser, as in central Ohio. Although the conditions of sedimentation are best described as quiet, the abundant sediment making up the member was probably swept to its present position and distributed largely by gentle current-action and not by subsidence from mechanical suspension in the water. The material is too coarse to be carried any distance in that way. The presence of such gentle current-action is shown, for example, (1) by frequent cross-bedding in the finest material with long, gently sweeping stratification planes, and (2) by the local sweeping together into a heap of such an odd aggregation of species and abundance of individuals as was found in three cubic feet in a certain bed at Sciotoville, which is otherwise practically barren of fossils, although excellently shown for half a mile and 30 feet thick.

If the correlations made by Cooper in sections to the northward be accepted, the Byer member continues little changed and from 25 to 40 feet thick for at least 55 or 60 miles to the northward of

central Ohio, into Wayne County.[1] This is to within 40 miles of Lake Erie.

The Allensville member.—The quiet conditions of Byer time were terminated by general shoaling of the region which resulted in the formation of the Allensville bed, whose chief characteristic is its coarse sandstones. Over much of the area this was brought about gradually, since there is a transition from one to the other; in fact, the typical sandstones of the Byer type are found almost to the top of the Allensville.

The name Allensville is given from numerous outcrops near that village in the western part of Vinton County, although the bed is not especially well developed there nor any better exposed than at other points. As is shown on the map (see Part I), it is found from the Ohio River as far north as the work has been pushed. In Scioto County along the river it is known certainly that it disappears westward, thus demonstrating that the source of its material lay to the eastward. In Pike and Ross counties this westward disappearance has not been completely demonstrated because of poor outcrops, but it is almost certain.

The greatest thickness yet observed is 39 feet, reported by Mr. Schroyer in Pike County, but this is exceptional. Commonly it is from 15 to 25 feet throughout the whole area but not infrequently it is less. It is least on the Ohio River near its extreme western limit where it diminishes rapidly from 6 feet to as many inches just before its final disappearance. The bed is apparently present at every point within the area of its occurrence indicated on the map.

As just noted, its chief characteristic is the occurrence of very coarse, excellently assorted, and usually well-rounded quartz grains. South of Fairfield County the material of the coarse beds is really a fine conglomerate of excellently assorted pebbles from one-sixteenth to one-eighth of an inch in diameter. The coarse beds there first appear in the section as thin layers, often only an inch or less in thickness, between much thicker beds of the typical, Byer-like sandstones. It is very difficult exactly to locate the base if the outcrops are poor. Much of the lower part of the member

[1] *Bull. Sci. Lab. Dennison Univ.*, V (1890), 24–32.

might well be considered the last stage of the Byer, but the lowest occurrence of the coarse beds has been adopted as the base of the member, even though this may not be the same particular streak that has been observed at the supposed base in a closely adjacent section. The coarse beds become more numerous and thicker in passing upward and usually make up the greater part of the upper half of the member. The topmost 3–5 feet of the member are almost invariably made up of these coarse sands to the almost complete or complete exclusion of the fine-grained material. At all points these coarse deposits carry unusually large amounts of iron, their cement being largely limonitic. This feature alone is often sufficient to identify the horizon. In southern Ohio, beyond the fact that the topmost bed is always the thickest, it is not possible to recognize the beds from section to section.

In Fairfield and Licking counties the material is noticeably finer. Whereas to the southward the beds may be described as made up of small pebbles, from Hocking County northward the component material can only be called coarse sand. The assortment is good and there is no notable admixture of the coarser material except in certain limited beds.

In eastern Fairfield County and central and eastern Licking County a characteristic feature of the member is the remarkable persistency of its beds. In its thickness of 15–25 feet, four distinct submembers can be recognized, all more or less persistent and with more or less distinct faunules. This condition is not found to the southward. Following is the section of the Allensville member as exposed in the Vogelmeir quarry at Newark; beds 1, 2, 4, and 5 can be recognized with considerable variation from Rushville to Newark, beyond which they have not been followed. Both Nos. 4 and 5 are likely to have lenses of coarse sand developed in them extensively and No. 5 at Rushville is largely of such coarse material.

	Ft.	In
5. Moderately fine-grained, irregularly bedded, yellow sandstones with numerous beds of coarse, reddish, iron-stained sandstone, the last mostly in the upper half, fossiliferous..............	14	
4. Bluish-gray, argillaceous shale, little grit, fossiliferous........	3	9

	Ft.	In.
3. Coarse quartz sandstone, grains well sized and rounded, white in color, harder and with much less iron than the other beds. This is Conglomerate II of Herrick and the bed which forms the top of the Black Hand in Prosser's description of this section......		½–6
2. *Allerisma winchelli* horizon, a hard, bluish, gritty shale with a considerable fauna, chiefly of lamellibranchs and abundant *Spirophyton* markings, one of the most important horizons in central Ohio..	5	
1. Very coarse, loosely coherent, reddish sandstone, grains uniformly sized and all about one millimeter in diameter, rounded or subangular. Lower surface slightly irregular..............	1	4

Evidence of erosion of the underlying beds before some of the coarse members were laid down is not unusual. Both at Rushville in Fairfield County and at Newark the topmost bed of the Byer sandstone had been eroded slightly before the basal bed of the Allensville was laid down. The fossils in the *Allerisma* shales are practically always found in the attitude which they maintained during life, the shells of *Prothyris* standing vertically with the anterior end downward, the shells of *Allerisma*, *Grammysia*, and *Sphenotus* standing obliquely with the anterior end downward. At Newark these shells have been observed at the top of the *Allerisma* shale with the upper portion beveled off at the contact with the overlying coarse bed (Conglomerate II). Evidence of erosion is also seen at the base of some of the higher coarse lenses when they are well developed. Obviously the erosion observed in connection with most of these beds is local. This urges caution in accepting the erosion which took place prior to the formation of the lowest bed as any more important. It is, however, probably of somewhat greater significance since in central Ohio it inaugurated an abrupt change in conditions of sedimentation. In the southern counties no evidence of such erosion has been seen, and the change in the conditions of sedimentation was more gradual. Among numerous interesting problems is the occurrence of these slight erosion surfaces associated with the finer beds of the member, whereas to the southward where the beds are coarser the evidence of erosion, if present at all, is less prominent.

Curious and probably of considerable significance is the manner in which the Allensville bed disappears from the column in southern

Ohio. At Slocum Station, $2\frac{1}{2}$ miles east of Sciotoville, the Allensville bed is well developed, with a thickness of 13 feet 4 inches. Of this the uppermost 4 feet 3 inches are composed almost wholly of the typical fine conglomerate which characterizes this member in southern Ohio. Three-fourths of a mile southwest of this locality only 2 inches of typical Allensville pebble material is present in a section sufficiently clear to show that no other beds of this type are present. A mile and a quarter west of Slocum Station only the merest trace of the Allensville member is present in the form of a few quartz pebbles of usual size in a 2-inch zone of fine-grained sandstone. At Sciotoville there is no evidence of its presence or suggestion of its proper horizon, although repeated search has been made for it in good sections. Not only are the pebbles wanting from this section, but there is no bed of coarse sandstone or bed of any description, which by any character suggests that it may be the westward extension of the Allensville bed. That member completely disappears from the Logan formation certainly within $3\frac{1}{2}$ miles and probably within 2 miles, with no tendency to grade into finer beds.

The Vinton member.—This is the youngest member of the Waverly series. How much more there may have been at one time is not known, since the Vinton is limited upward at all points, where it has not been entirely removed, by an erosion surface, a horizon of great unconformity.

The Vinton member is named from Vinton County, where it is well developed, although not nearly so well developed as elsewhere. It is a series of fine-grained gray or yellowish sandstones and shales. It differs from the typical Byer in the presence of shale beds and in the thinner bedding and more platy nature of the somewhat harder sandstones. However, as noted under the Byer member, these members come to resemble each other very closely in southern Ohio.

Although no especial effort has been made to trace them, it is very evident that certain of the beds composing this member are quite persistent areally. This not only indicates the uniformity of conditions which prevailed but may have yet a greater significance, since there is considerable faunal difference in certain of the

different beds. For example, at the base of the formation from central Licking County southward at least as far as western Vinton County, a distance of 55 miles, there is a 30- to 40-foot bed of sandy shales with no particular faunal characters noted as yet. It is overlain at all points by a bed of fine-grained yellow sandstones with an abundant fauna, of which one of the chief species is a *Productus*, heretofore described as, but distinct from, *Productus arcuatus* Hall.

The Vinton member marks the resumption of conditions of quiet, uneventful deposition in deeper water, following the period of shoaling which resulted in the formation of the Allensville member. There is no transition from the one to the other. The shales of the former rest abruptly on the coarse beds of the latter at all points, and this contact is one of the sharp and persistent ones in the Waverly series. It is suspected that its significance as a stratigraphic horizon is more or less in proportion to this prominence.

The thickness of the Vinton is determined wholly by the attitude of the post-Waverlian erosion plane. Valleys of considerable depth were cut into the Waverly following its deposition and prior to the accumulation of the Pennsylvanian deposits, or Coal Measures, and considerable irregularity can be detected in its upper surface as a result. The greatest known thickness of the member is preserved in the northwest portion of Perry County and adjacent parts of southern Licking County. Here was an area which must have been uniformly high land, for the Vinton member is from 200 to 240 feet thick, although it does not usually much exceed 50–75 feet thick, and is occasionally wholly removed. On the Ohio River between Sciotoville and Portsmouth the member probably attains a similar thickness, but precise determination is seriously hampered by the absence of the Allensville member. The Mississippian-Pennsylvanian contact in central and southern Ohio has been briefly described.[1]

[1] J. E. Hyde, "Notes on the Absence of a Soil Bed at the Base of the Pennsylvanian of Southern Ohio," *Am. Jour. Sci.*, 4th ser., XXXI (1911), 557–60.

REVISION OF THE MAP OF LAKE AGASSIZ

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After the lapse of twenty years since the completion of my official service on the geological surveys of Minnesota and the United States, it is not surprising that changes and additions have been made, and that others are needed, on maps of the glacial and modified drift formations and the successive shorelines of the glacial Lake Agassiz, as these were published in the reports of my work from 1879 to 1895. Therefore the recent *Bulletin 12* of the Minnesota Geological Survey,[1] entitled "Surface Formations and Agricultural Conditions of Northwestern Minnesota," by Mr. Frank Leverett, of the United States Geological Survey, has much interested me.

One of the most noteworthy additions here made to our knowledge consists in the more accurate description and mapping of the somewhat elevated large tract between Red Lake and the Lake of the Woods, which was named by me Beltrami Island.[2] My original description and map of this island were doubted and questioned by Professor James E. Todd in 1899;[3] and they have now been shown by Mr. Leverett, with more definite surveys, including information of exact altitudes throughout this area, to need important revision.

A line surveyed about the year 1892 for the Duluth & Winnipeg Railroad Company took a straight northwesterly course from the most northeastern part of the shore of Red Lake to the most south-

[1] This report, 78 pages, with 8 plates (a folded map and views from photographs) and 14 figures in the text (maps and diagrams), was published by the University of Minnesota in February, 1915, under Professor William H. Emmons, director, in co-operation with the United States Geological Survey. It includes also a chapter on "Climatic Conditions in Minnesota," by U. G. Purssell, United States weather observer in Minneapolis.

[2] *Am. Geologist*, XI (June, 1893), 423–25; "The Glacial Lake Agassiz," 1896, *U.S. Geol. Survey Mon. XXV*, 304–06, with mapping on Plates III, X, etc.

[3] *Geology of Minnesota*, IV, 149–50.

western part of the Lake of the Woods, at the present station of Warroad. Through a distance of 18 miles next northward from Red Lake this survey traversed a continuous swamp, called on the profile "tamarack and spruce swamp," which includes the divide between the basins of this lake and the Rainy River. The depth of the mud and water throughout the swamp was from 5 to 15 ft., being mostly between 7 and 10 ft.; and in many places the hard bottom was noted as gravel and sand. The highest part of the surface of the bog or swamp on this line, forming the water divide, some 5 to 6 miles north of Red Lake, has an elevation of 1,197 ft. above the sea, being 21 ft. above the lake. Westward from this lowest part of the divide on the north side of the basin of Red Lake and the Red Lake River lies the area of Beltrami Island, of which the northeast margin was crossed by this railroad survey, rising there to a height of 1,283 ft., or 107 ft. above Red Lake and 86 ft. above the eastern base of the island.

The profile of the surveyed line shows a well-defined beach of Lake Agassiz at 1,215 ft., this being the highest recognizable beach ridge crossed by the survey. On the descent toward the Lake of the Woods the profile crosses a succession of ten lower beaches of the great glacial lake, their altitudes above the sea being 1,196, 1,172, 1,156, 1,143, 1,127, 1,116, 1,106, 1,099, 1,093, and 1,087 ft.[1]

These beach ridges are clearly indicated by the profile, but none are shown at higher levels. My conclusion from the profile record, therefore, was that the series comprises the highest shore and the successively lower shores of Lake Agassiz. The higher part of the tract northwest of Red Lake was accordingly called an island of that glacial lake, and was named, like the county in which it mostly lies, for the exiled Italian author, Costantino Beltrami, who in 1823 was one of the earliest white men to traverse and describe this region. He was preceded only by the Canadian geographer, David Thompson, in 1798.

An elaborate map of this area, from Red Lake north to Rainy River and the Lake of the Woods, with contour lines at intervals of 10 ft., accompanies a report published in 1909,[2] entitled "Drain-

[1] Notes of this profile were published in *Geology of Minnesota*, IV (1899), 136–37.

[2] 61st Cong., 1st sess., *H. R. Doc. No. 27*, 40 pages.

age Survey of Certain Lands in Minnesota." These lands, ceded by the Chippewas, were surveyed in 1906–8, for the United States Geological Survey, under direction of W. H. Herron, geographer. On the large map of this report the contour of 1,200 ft. above the sea incloses a tract of 1,167 square miles, its most southern point being 4 miles north of the mouth of Red Lake, and its most nothern point 2 miles south of Roosevelt, on the Canadian Northern Railway, and about 9 miles south of Muskeg Bay, the most southwestern part of the Lake of the Woods. This large tract, nearly flat or somewhat undulating, rising to a maximum elevation of 1,310 ft., would be an island if the conditions that produced Lake Agassiz were now restored with the lake level at 1,200 ft.

Beltrami Island, as thus outlined, however, differs much from its mapping in the Lake Agassiz monograph; for on the west it falls short, by 12 miles, from reaching to Thief Lake, which had been included in the tentative original map of the island area. But the beaches bordering the island have been only partially mapped; and when this work shall be completed by leveling they will show an ascent northward, or almost northeastward, like the other mapped shores of the old glacial lake. Appealing to the mapped gradients of the Herman shorelines from the south end of Lake Agassiz north to Maple Lake and eastward past the south side of Red Lake, Mr. Leverett well concludes that the highest Herman beach, if it is represented on Beltrami Island, must sweep with its northeastward ascent well-nigh above the entire island, touching only its highest ground, in Tp. 159, Rs. 35 and 36.

Even the second, third, and lower Herman beaches, mapped in succession at slightly lower lines to an entire number of seven on the west side of Lake Agassiz in southern Manitoba, may all be referable, as Professor Todd suggested, to water planes above the beaches at 1,215 and 1,196 ft. on the railway profile crossing the northeast edge of Beltrami Island. My original supposition that much of this island was above the highest level of Lake Agassiz has been needfully corrected by Mr. Leverett in this report.

Whether any of the Herman shores was marked on the island must depend on the position of the ice border in northwestern Minnesota when the glacial lake fell to its Norcross stage, which,

like the Herman stage, and like the lower Tintah, Campbell, and McCauleyville stages, becomes subdivided in its extension northward before reaching the International Boundary at the west side of Lake Agassiz. In my examination and studies of these shorelines for the geological surveys of the United States and Canada, the whole series of seven Herman beaches was regarded as yet present so far north as on the boundary between North Dakota and Manitoba, 224 miles north of Lake Traverse; but the two upper Herman beaches were thought to cease before coming to the latitude of Gladstone and Neepawa, in Manitoba, 308 miles from Lake Traverse and the mouth of the glacial lake. On Valley River in Manitoba, between the Riding and Duck mountains, 150 miles north of the international boundary and 375 miles from Lake Traverse, all the Herman beaches had ceased, according to my tabulation and correlation of the notes of Mr. J. B. Tyrrell, of the Canadian Geological Survey; but the Norcross, Tintah, Campbell, and McCauleyville beaches seem to be represented there by his observations.[1]

The discrepancy between my estimates and those of Mr. Leverett for the place or height of the Herman beaches north of Red Lake comes partly from his better determination of the large easterly component in the direction of maximum uplift and present ascent of the ancient lake-levels, which in their continuation eastward of Maple Lake trends increasingly toward the northeast. In manuscript notes on a map kindly supplied to me, Mr. Leverett records a well-defined beach of Lake Agassiz, 1,297–1,300 ft. above the sea, on the Minnesota & International Railway about 4 miles northeast of Gemmell, or 25 miles east-southeast from the north part of Red Lake. This is the highest of the Herman series there, and its rate of ascent from the vicinity of Maple Lake justifies the belief that has been already stated concerning the occurrence of the Herman shorelines on the area of Beltrami Island.

It is to be noted further in this connection that the trend of the Late Glacial uplift of the land and consequent ascent of the old beaches, near Red Lake and the Rainy River, now found to be N. 35° E., carry the conspicuous Campbell beaches farther east

[1] *U.S. Geol. Survey Mon. XXV*, pp. 476–77.

than my mapping in the United States Geological Survey monograph. The incoming of settlers, with the construction of roads and railways and clearing away the forest in many places to open farms, now gives opportunity to trace the courses of the beaches through the country adjoining the Rainy River, where Mr. Leverett maps a finely developed shoreline, referred to the Campbell stage of Lake Agassiz, for the distance of about 40 miles, from near Roosevelt east-southeastward to Baldus, having the height along all this isobasic course of about 1,120 ft. above the sea. This altitude is as great as the highest of the Campbell beaches has on the eastern slope of Riding Mountain in Manitoba, which shows that the isobase of the land uplift there crossing the lake area runs about N. 55° W. In other words, the direction of maximum uplift is approximately N. 35° E.

With the light thus received, we may quite confidently correlate the series of eleven beach ridges noted by the profile on Beltrami Island and northward, from 1,215 to 1,087 ft., as representing the second Norcross, Tintah, and Campbell stages of Lake Agassiz, but with other subdivisions than were noted in my tabulation for the latitudes of Gladstone and Valley River, Manitoba.

The present report and its large detailed map of the surface formations of the northwest quarter of Minnesota, on the scale of 8 miles to an inch, are designed primarily to instruct and aid farmers, drainage engineers, road-builders, and others, for the development and utilization of the agricultural resources of this state.

Glacialists and other geologists will desire that Mr. Leverett present later, for the entire state of Minnesota, a more specific description and discussion, and additional maps, setting forth the successive borders of the waning ice-sheet, as shown by their moraines, with the corresponding stages of growth of Lake Agassiz, and treating fully of the erosion, transportation, and deposition of the various drift formations. Such maps will be much improved by insertion of contour lines, and by notations of altitudes of lakes, rivers, railway stations, etc., whereby the form of the land surface and the elevation above the sea will be shown, giving information that seems indispensable for a definite understanding of the drift, loess, and glacial lakes.

Reviews

Geology of the Nome and Grand Central Quadrangles, Alaska. By FRED H. MOFFIT. Bull. U.S. Geol. Surv. No. 533, 1913. Pp. 136.

The Nome and Grand Central quadrangles are situated in the south-central part of Seward Peninsula in an area included between 165° and 165°30′ west longitude and 64°25′ and 64°57′ north latitude. The Tigaraha schist which is the uppermost member of the Kigluaik group (early Paleozoic and possibly in part pre-Paleozoic) is the oldest formation exposed. One principal granite mass and smaller dikes and sills intrude this schist. The relation between the Tigaraha schist and the overlying Nome group appears to be one of conformity. The Nome group (probably middle Paleozoic) makes up the greater part of the mapped area. It consists of two schist horizons with an intervening limestone formation. These beds are cut by greenstone dikes and sills and to a small extent by granite. Quaternary (possibly including some Tertiary) deposits that consist of marine and fluvial sands and gravels and glacial débris succeed the Nome group. The mineral wealth of the Nome district lies in its placer deposits. The richest of these are rapidly decreasing, so that the present production can be maintained only by new and better methods. The placers are described as residual, stream, beach, and gravel plain. Dredging will play an important rôle in the future development. A historic outline of the Nome placer development is given. There are gold, stibnite, bismuth, scheelite, copper, and graphite lodes, but these are at present not of commercial importance.

V. O. T.

Coastal Glaciers of Prince William Sound and Kenai Peninsula, Alaska. By U. S. GRANT and D. F. HIGGINS. Bull. U.S. Geol. Surv. No. 526, 1913. Pp. 72.

The aim of this report is to "supply some definite information regarding the present positions of the fronts of the glaciers" of Prince William Sound and Kenai Peninsula and "the more evident facts of their fluctuations." The Barry, Surprise, Chenega, Princeton, and Holgate glaciers have retreated from one to two miles within from 10 to 50 years or more. Since the growth of the present coniferous forest, the Columbia and Bainbridge glaciers have recently advanced to their maximum positions. The glaciers on the west side of College Fiord and Harvard Glacier have been advancing for the past ten years; their maximum advance was in 1905. Many of the glaciers have exhibited alternate advances and retreats of notable extent. There is no indication of a general advance or retreat during the last 50 years. These glaciers are remnants of an ice sheet whose upper surface once reached an elevation of 2,000 feet at the main coast line. The accompanying photographs (that constitute 40 plates) of the glaciers are exceedingly good and instructive.

V. O. T.

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THE LATER STAGES OF THE EVOLUTION OF THE IGNEOUS ROCKS

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CONTENTS

INTRODUCTION

With the recent appearance of the books of Harker, Iddings, Daly, and von Wolff, it is natural to expect that the subject of our present knowledge of the igneous rocks and their origin has been thoroughly stated. Nevertheless, it has seemed to the writer that in these treatments of the subject, largely from the point of view of the field geologist, some questions have been left open which are capable of definite solution if the bearing of recent experimental work on these subjects is appreciated. Much of the experimental material was, indeed, not available when these books were in preparation. An attack on the problems of the igneous rocks from this latter point of view has led to a conception of their origin which is stated in the present paper. It is not hoped that the brief statement offered can possibly be the whole truth or that erroneous conclusions have been entirely avoided in a subject of such magnitude and complexity. Some hope is entertained, however, that the discussion offered may have a tendency to turn petrologic thought in a definite direction, and, perhaps, may suggest subjects of experimental investigation which may be pursued with likelihood of finding useful application of their results.

Differences in composition in associated rock bodies and in various parts of a single body have been referred by petrologists to two principal processes: assimilation of foreign material and differentiation. According to individual opinion, one or the other has been favored and some have proposed processes involving a combination of the two. A consideration of field facts, especially

in the light of the results hitherto obtained in the experimental study of silicate mixtures, has led the writer to believe that the former process, assimilation, is comparatively unimportant quantitatively, and that to understand the igneous rocks we must direct our energies toward the elucidation of processes of differentiation. Moreover, of the various processes of differentiation that have been proposed, it is considered that differentiation, or fractionation, by crystallization is vastly the most important. In the present paper the facts which have led to these conclusions are stated and a systematic petrogenic theory based on these conclusions is proposed.

DIFFERENTIATION

The term differentiation is applied to any process whereby a magma, without foreign contamination, forms either a mass of rock that has different compositions in different parts or separate masses that differ from one another in composition. The subject has been much discussed and apparently every conceivable process given a place as a factor in producing the observed results. In the following pages, discussion of these processes will be entered into only in so far as is necessary in order to indicate the reasons for believing that but one of these is of great importance, namely, crystallization and the movement of crystals relative to the remaining liquid.

Some investigators have considered that difference of composition in the magma itself, due to a temperature or a pressure gradient or to gravity, is the important factor. These continuous variations in the composition of the liquid, supported by Vogt, Iddings, and others, which are due to a *Diffusion oder Wanderung*[1] in a single liquid phase,[2] are to be carefully distinguished from a breaking up of the magma into immiscible partial magmas, i.e., a separation into two or more liquid phases, a process supported by Rosenbusch, Bäckström, and others.

Theoretically, continuous composition gradients should be established if sufficient time were available. However, in an actual case, there is reason to believe that crystallization begins long

[1] J. H. L. Vogt, "Über anchi-monomineralische und anchi-eutektische Eruptivgesteine," *Vidensk. Selsk. Skr.*, I Math.-Naturv. Kl., 1908, No. 10, p. 6.

[2] When a phase shows a composition or concentration gradient it can be treated in phase-rule considerations as divided into "regions"; see R. C. Tolman, *Jour. Am. Chem. Soc.*, XXXV (1913), 307.

before even very small effects due to these causes can be obtained. With crystallization, possibilities of differentiation are introduced which completely overshadow any of the possible effects of diffusion in a liquid phase.

THE SORET ACTION

The kind of composition gradient most commonly considered is that due to a temperature gradient, commonly known as the Soret action. It is generally assumed that a concentration takes place toward the colder parts of those substances whose crystallization is imminent. This assumption is designed to explain the observed fact that there is often a richness in the minerals of early crystallization toward the boundaries of igneous-rock bodies. But the Soret action takes place entirely in the liquid and is wholly independent of the solubility of the various substances in the liquid. There is therefore no reason to believe that the substances concentrated in the cold part would be those that are about to separate out. If a tube filled with a very dilute solution of common salt is warmed at one end, there will result a concentration of salt toward the cold end. If the whole tube is now gradually cooled, this same difference of temperature being maintained, a time will come when ice, not salt, crystallizes at the cool end. The substance concentrated at the cool end as a result of the Soret action was salt, the substance first crystallized there was ice. The Soret action has produced an effect in this case precisely the opposite of that which is commonly supposed to result from it. It is clearly unsafe to assume that the substances concentrated toward the cooling boundary would be those of early crystallization.

There is also a difficulty in obtaining a sufficient quantitative effect from the Soret action. Harker has demonstrated this truth by imagining certain temperature differences and calculating the possible composition differences that could be produced,[1] assuming that the concentration varies as the absolute temperature. This calculation shows that for a temperature difference of 100° at 1200° the relative concentrations would be as 1.07:1.[2] Even this esti-

[1] *The Natural History of Igneous Rocks*, p. 316.

[2] The Soret action has recently been reinvestigated for salt solutions and it has been found, as might be expected, that the simple relation deduced from osmotic-pressure laws does not hold (August Eilert, *Zs. f. anorg. Chemie*, LXXXVIII [1914], 1).

mate sets rather an upper limit which is probably many times the actual effect, for nearly all experimental investigations of the Soret phenomenon agree in placing the actual effect at much less than the amount calculated on the foregoing assumption.

The establishment of this fact should not be surprising, for the Soret effect does not, in reality, depend on a tendency of the solute alone to become more concentrated in the cold part. There is a tendency toward a greater concentration there of both solute and solvent, which is of course impossible, and if both obeyed the laws of ideal solutions, there could be no relative concentration of the one with respect to the other.[1] It is because the behavior of solute and solvent departs from the laws of ideal solutions, the one more than the other, that the Soret phenomenon is observed. The amount of the effect obtained as a result of this difference in degree of departure from the laws mentioned must necessarily be less than that calculated from the osmotic-pressure relation, which really involves the assumption that one substance obeys these laws and the other substance may be neglected, that is, is entirely unaffected by change of temperature.

Both theoretical considerations and experimental results indicate, therefore, that the actual Soret effect would be even less important than Harker's calculation would indicate—in short, quite negligible.

GRAVITATIVE EFFECT IN A LIQUID MAGMA

Of the same order of magnitude is the probable effect of the accumulation of the denser components of the magma solution in its lower portions, resulting in a composition gradient or density stratification in the liquid magma. An experiment made by Morozewicz has sometimes been cited as a proof that the extent to which this process may take place even in a crucible is very considerable, but a consideration of the conditions of the experiment shows that this conclusion is unwarranted.[2] In the experiment referred to, a sample of granite was fused or rather partly fused and, after cooling, analyses were made of the glass from the top and from

[1] See Bäckström, *Jour. Geol.*, I (1893), 774–75.

[2] J. Morozewicz, *Tscherm. Min. Pet. Mitth.*, XVIII (1898), 232.

the bottom of the crucible. The analyses showed a considerably larger percentage of silica in the upper portion, but it was also noted that there were grains of undissolved silica floating in the upper part. With silica in process of solution near the top, it would be a marvel indeed if the liquid in the upper part was not more siliceous than that in the lower, when the extreme viscosity of the liquid, and the consequent impossibility of diffusion and convection of sufficient vigor to maintain uniformity, are considered. Morozewicz' figures show also that there was a greater concentration in the lower portion, as compared with the upper portion, of iron oxide and the alkaline-earth oxides and a lesser concentration of the alkalis. The natural explanation of this is that some of the alkaline feldspar collected in the upper portion in exactly the same manner as the grains of quartz. After its solution, which for the alkaline feldspar happened to be complete, apparently, diffusion and convection were unable to establish uniformity of composition in the liquid. It should be noted, moreover, that the percentage of iron oxide and of lime was greater in *both* the lower and the upper parts than in the original granite. We must, therefore, add to the complex conditions pointed out above a contamination from the pot or furnace, at least from some outside source. Clearly no significance can be attached to the experiment in the matter of indicating the effect of gravity in producing composition differences in a liquid.

The rather rapid melting of a mixture of minerals such as a granite cannot be a simple process. It begins with the liquefaction of those substances which flux most readily together, not necessarily those of lowest melting-point. The substances first melted are principally those rich in alkaline earths and iron. The liquid, as gradually formed, tends to filter down through the porous material and collect toward the bottom. Thus there results in the early stages a relative concentration of alkalic and siliceous materials toward the top. The more siliceous and alkalic substances, which assume the liquid state with great reluctance, dissolve slowly, and in time all becomes liquid. The melt will, however, be layered as a result of the accumulation of the first-formed liquid at the bottom and possibly the floating of crystal grains buoyed up by

gas bubbles at a later stage. The persistence of this layering, or, better, continuous composition gradient, is due only to the inability of diffusion to overcome it in such viscous melts in any reasonable time. The phenomenon was noted long ago by Sorby. In describing the melting of Mount Sorrell "syenite," a hornblende granite, he says, "The hornblende melted more easily than the quartz and feldspar, and a portion of those minerals has risen upward."[1]

The phenomena which take place in the glass-maker's pot are essentially of the same general nature; that is, the difficulty connected with obtaining a homogeneous melt in glass practice is due to initial inhomogeneity and not to any tendency of the liquid, once homogeneous, to become inhomogeneous. The truth of this statement becomes obvious when it is realized that the glass-maker, in order to reduce inhomogeneity, resorts to long-continued heating, thereby affording diffusion sufficient time to accomplish the desired result. If there were any tendency to *become* inhomogeneous, continued heating would give precisely the opposite effect.

SEPARATION OF DISTINCT PHASES

The possibilities of composition differences that have been considered are those which may conceivably exist in a magma still entirely fluid and consisting of but one phase. They are not adequate to explain the differences actually observed in nature in the crystallized product of the magma. As soon, however, as new phases begin to separate from the magma, the possibility of obtaining composition differences in different parts is enormously increased, if there is any mechanism whereby a relative concentration of the new phase or phases in certain parts may be accomplished. The new phases might be either liquid or crystalline and the factors which might result in their concentration in certain parts are diffusion, convection, gravity, and earth movements. Consideration will first be given to the question whether the separation of liquid phases, i.e., immiscibility in the liquid state, is a factor in differentiation.

[1] *Proc. Geol. and Polytech. Soc. W. Yorkshire,* IV (1863), 302.

Liquid immiscibility in magmas.[1]—It is well known that many liquids which are quite homogeneous at high temperatures may separate at lower temperatures into two or more non-consolute liquid fractions, and the hypothesis that igneous magmas form such immiscible fractions is favored by some investigators. Often it seems not to be realized that the formation of an immiscible liquid fraction is not in itself sufficient for differentiation. There must be some mechanism for the relative concentration of the immiscible liquid in certain parts, and those who deny the efficacy of, say, gravity in the production of a local concentration of certain crystalline phases must, logically, do the same for immiscible liquid phases.

Vogt has made a study of the question of liquid immiscibility in silicates, considering both experimental and field evidence, and has arrived at the conclusion that the rock-forming silicates are freely miscible in all proportions. For certain minerals like the sulphides, he concedes partial immiscibility.[2]

Some geologists are inclined to disagree. Daly, for example, in speaking of the complete miscibility of liquid silicates, says: "It may be quite true for high temperatures and yet quite untrue for a temperature just above that of crystallization of a given component. No one has yet succeeded in holding a molten mixture of silicates within this narrow range of temperature for a length of time sufficient to warrant any conclusion on the matter."[3] On the contrary, in the great number of quenching experiments which have been made at the Geophysical Laboratory on a rather compre-

[1] The term "liquation" is now quite commonly used in petrologic literature to designate the act of separation into immiscible liquid fractions. This modern usage has probably been the cause of the insertion of this meaning in the supplement to the *Century Dictionary.* Schweig appears to apply it, however, to the process of separation of liquid from crystals which he contrasts with separations of any kind in the liquid, a contrast always to be carefully borne in mind in any discussion of differentiation. Schweig's usage appears to be the more justifiable on the basis of the borrowing of the term from the metal-worker, to whom it indicates a process of separation of a liquid portion from a crystalline or solid portion such as that accomplished in the Pattinson process for the purification of silver. However this may be, the writer considers it advisable to avoid the term, since, without its context, it does not convey a definite meaning especially with regard to the very much contrasted processes mentioned above.

[2] *Silikatschmelzlösungen*, I, 96; II, 228.

[3] Daly, *Igneous Rocks and Their Origin*, p. 226.

hensive variety of silicate mixtures, ideal conditions are maintained for the discovery of liquid immiscibility. The charges are held at various temperatures, both above the temperature of beginning of crystallization and throughout its range, for a period of time sufficient to obtain equilibrium between the liquid phase and the crystalline phases. How much more readily would equilibrium have been obtained between two liquid phases if there had been any tendency toward the formation of two such phases! Yet not a single instance of liquid immiscibility has been encountered in the whole course of this work. The experiments have been conducted both at atmospheric pressure and in some cases in bombs under high pressure of water-vapor.

It may be pointed out that, for the detection of immiscibility, it would not have been necessary to allow sufficient time for the collection of the immiscible portions into separate layers. The separation of globules of liquid 0.01 mm. in diameter and even much smaller could not have escaped detection during the microscopic examination of the quenched products. Continued experimental work is, then, in accord with Vogt's conclusion, and it is safe to repeat that the experimental evidence, as far as it goes, is decidedly against liquid immiscibility.

In the case of natural rocks, differences of composition in various parts of a body, when completely crystalline, might be referred equally well to the separation either of crystalline or of liquid phases. Similarly, there is no compelling reason for preferring liquid immiscibility as the explanation of the association, however intimate, of *distinct* lava flows of different composition, such as the basaltic and rhyolitic lavas of Iceland. Granting that both came from the same source, it is possible that the extrusion of the different varieties was separated by periods of years, during which interval there was plenty of opportunity for a change in the composition of the liquid in the magma basin due to the separation of crystals from it.

Even the cases, sometimes described, of simultaneous intrusion or extrusion at the same locality of different magmas does not prove their coexistence as *immiscible liquids* in the same pool. There is often, on the contrary, distinct evidence, when two magmas are

brought together in this manner, that they have a strong disposition to mix and do so to the extent of their ability before crystallization or cooling to the glassy state ensues. The facts point rather to derivation of these magmas in most cases from separate, though adjacent, pools, say, laccolithic chambers.[1] The pools may originally have contained the same magma, but on account of a difference in size and consequently in rate of cooling were, at the time of reintrusion, at different stages in their career of crystallization and therefore, as will be shown in the sequel, the still liquid portion drained-off from the one had a different composition from that drained from the other.

There is only one kind of phenomenon which could be considered as definite proof of the occurrence of limited liquid miscibility and it would, without doubt, be very commonly observed if limited miscibility among rock-forming silicates were a fact. The observed phenomenon would be the occurrence in glassy, or partly glassy, extrusive lavas of distinct globules of material still partly glass, large or small according to their opportunity for aggregation, and of composition different from that of the main mass. Here would lie indisputable proof of the formation of immiscible liquid globules. So far as the writer is aware, no such case has been described, in spite of the fact that lavas, as a group, depict all igneous processes arrested, more or less successfully, at all possible stages. They are Nature's quenching experiments.

The rare globular masses found in those completely crystalline plutonic rocks to which the term orbicular has been applied are an entirely different matter. It is often demonstrable that these owe their origin to the localization of *crystallization* of certain constituents, due to the cooling and chemical effect of a foreign inclusion, the remains of which often constitute the nucleus of the orbule.[2] The occurrence of such structures as an exceedingly rare curiosity shows that they always require special conditions and are not a normal result of cooling as the formation of immiscible

[1] Certain stages of crystallization at which the liquid portion may have different compositions in different parts of a single reservoir will be discussed later.

[2] Von Chrustschoff lists other possibilities (*Mém. Acad. Sci. St. Pétersbourg* [7], XLII, No. 3).

liquid portions should be. If they are ever due to a separation in the liquid state, it is remarkable that sudden extrusion has not happened to give plain evidence of the fact in the occurrence of similar orbules of *undercooled liquid* in the extrusive types. In the absence of evidence of this kind, it is necessary to reject liquid immiscibility as a factor in the differentiation of igneous rocks. Lavas could not fail to show the separation of liquid phases in all its stages just as clearly as they do the separation of crystalline phases in all its stages. Attention must be directed toward the separation of crystalline phases—i.e., crystallization—as the fundamental factor in differentiation though the possibility that sulphides and perhaps some oxides may become immiscible under certain conditions may be freely admitted. It may be pointed out that, considered on general grounds, limited miscibility between such dissimilar substances as sulphides and silicates is likely, but wholly unlikely among the silicates themselves, substances which have such remarkable tendencies to be miscible even in the solid state.

Differentiation through crystallization.—Differentiation may be brought about as a result of crystallization in two ways: through the localization of crystallization and through the local collection of crystals.

Since the outer parts of a body of magma are always cooler than the interior, there is a certain period during which crystallization is taking place only near the border, the interior being still above the temperature of beginning of crystallization. It has been supposed that during this period precipitation of the mineral or minerals of early crystallization takes place in the cooler parts, but that this does not bring about a local impoverishment of the liquid in the crystallizing material, the composition of the liquid being kept uniform and continued growth of the crystals being maintained by free diffusion from all parts of the liquid[1] or by convection currents.[2] Thus there results a general richness of the border phase in minerals of early crystallization, according to this belief. Such freedom of diffusion Becker has shown to be altogether inconsistent with what is known on that subject. Apart

[1] A. Harker, *The Natural History of Igneous Rocks*, p. 318.

[2] G. F. Becker, *Am. Jour. Sci.* (3), 1897, pp. 21–40.

from this the rocks themselves show that diffusion is quite limited in its scope. When centers of crystallization are once established at the border, diffusion as free as that postulated should lead only to the continued growth of the crystals so initiated. There is no reason why new centers should be established. The result should be the growth inward from the contact of huge crystals of the minerals of early separation. In reality, of course, a new center of crystallization for these minerals is always established within a few millimeters, a fact which speaks eloquently for the radius through which diffusion acts freely.

The same objections apply to the supposed maintenance of approximate uniformity in the composition of the liquid through the agency of convection currents, or, better stated, perhaps, the continual supply of new liquid to the border portion. The same huge crystals at the border should result.

Let us suppose, however, that, in some accidental manner, say, before convection has fairly started, many centers of crystallization of these early minerals are established near the border. If these centers were fed by convection currents coming from distant parts of the magma, none of the crystals except those actually attached to the wall rock would remain in the border portion. They would be carried away and distributed by the same currents that are supposed to feed their growth. If the numerous early crystals established near the border are supposed to grow by diffusion of material from a distance, there is a similar difficulty. Such a process would require a long period of time and there is no reason to believe that the crystals would remain there when the border concerned is, say, the upper margin of a sill. The crystals would in this case sink out of the border portion.[1]

[1] Experimental work forbids the belief that diffusion could take place freely *through considerable distances* in a period of time too brief to permit significant sinking of crystals (cf. Harker, *op. cit.*, p. 322). For example, a liquid consisting of 56 per cent $CaMgSi_2O_6$ and 44 per cent $MgSiO_3$ permits quite definite effects in the way of sinking crystals at 1430° in only 15 minutes, but if one tries to make the foregoing liquid by carelessly mixing diopside and $MgSiO_3$ and holding the mixture at a temperature considerably above 1430°, little headway toward a homogeneous melt will be made in 15 minutes. The materials must be very carefully mixed in the form of a fine powder (N. L. Bowen, *Am. Jour. Sci.* [4], XXXIX [1915], 176).

If either free diffusion or convection toward the cool boundary were the effective agent in producing the concentration in that region of the minerals of early crystallization, we should expect large crystals of these minerals attached directly to the wall rock, or if the crystals were of small size they should be attached to each other and ultimately to the wall rock. In nearly all cases these early minerals occur, on the contrary, in small discrete crystals separated from each other by crystals of later formation which must have been a part of the liquid magma at the time the supposed diffusive or convective processes were taking place. At this time, then, they could not have been attached either directly or indirectly to the wall rock and there is no reason why they should have remained in the border position.

For the reasons above outlined, the processes of diffusion or convection during crystallization cannot in general be considered satisfactory explanations of "basic border phases." Diffusion toward the cold boundary prior to crystallization, the Soret action, has also been found wanting. Daly[1] has offered an explanation of basic border phases which seems to be entirely satisfactory in most cases. He pictures the border phase as a quickly chilled portion having the composition of the original magma and the more salic phase adjacent to it as the lighter differentiate formed during the slow cooling of the rest of the magma.[2] For stocks, batholiths, and large laccoliths, this process has distinct advantages over a process involving diffusion toward a chilled border. Daly has in this manner successfully explained that classic example of a body exhibiting a basic border, the Shonkin Sag laccolith.

In the case of small bodies like most dykes, it seems to the writer to be more commonly true, for those having basic border phases, that the dyke fissure serves as a channel for the passage of magma during a considerable period. As the chilled layer forming on the cold walls gradually increases in thickness, the composition of the magma passing through gradually becomes more acid on account of a like change in the composition of the magma in the

[1] Daly, *Igneous Rocks and Their Origin*, p. 244, Fig. 123, and p. 246, Fig. 126.

[2] In a later part of the paper the manner in which quick cooling prevents differentiation and slow cooling promotes it will be shown in detail.

feeding reservoir as crystallization proceeds. Some composite dykes at Cape Ann, on the Massachusetts coast, give distinct evidence of this process. These dykes are more salic in the central portion, and in places there is a continuous gradation in composition toward the more basic margin. Here the variation might be referred to differentiation in place, but that this conclusion would be erroneous is shown by the fact that in other places in the same dyke the more basic phase has been completely frozen, and only after a reopening of the dyke by fracture has the more salic phase entered. This refracturing has in some places been repeated. The evidence is clear that the dyke fissure has served for the passage, throughout a considerable period of time, of material of continually changing composition.

In general this explanation of basic borders when they are shown by small bodies such as narrow dykes or sills seems preferable to any process of differentiation in place.

The most promising processes for the production of differentiation are those involving crystallization and the relative movement of the crystals with respect to the liquid from which they separated. Of these the most important during the period when the magma is still dominantly liquid is the movement of crystals in the liquid on account of differences in density, usually the sinking of crystals. During the period when the magma has largely crystallized, "the straining-off or squeezing-out of the residual fluid magma"[1] is the most important process. It is to be noted, however, that in the strained-off, reintruded liquid from the latter process, the sinking of crystals may begin again, so that it is difficult to refer either of these processes to any definite period in the complete history of any given magma.

No one can reasonably question the existence or the effectiveness of a process of squeezing out of residual magma, but many have maintained that there is no evidence that crystals sink in magmas. Numerous examples, however, have now been described of sheetlike bodies with both roof and floor exposed and the differentiates arranged in such a way as to point clearly to the sinking of crystals. An antagonistic view is, therefore, no longer tenable. Those

[1] A. Harker, *op. cit.*, p. 323.

opposed to the process seem to expect too much of it, especially in the way of a sorting of the individual minerals. Thus it is stated that the iron ores do not accumulate at the bottom of the sill. This statement is not strictly true, for sometimes they do;[1] nevertheless such a phenomenon is to be expected only rarely. Let us assume the most favorable case. The separation of the ore minerals as the first products of crystallization takes place and their separation is complete before other crystals appear, a thoroughly unlikely assumption. The period of crystallization of the ore minerals *alone* would be even in this case exceedingly brief for bodies of moderate dimensions, and such are most bodies whose floors are visible. It is to be noted, moreover, that the very heavy minerals usually occur as relatively small crystals and the size of crystals is a factor of prime importance in determining their rate of sinking. Very small crystals of magnetite would, as a rule, make little headway before crystals of other minerals were precipitated.[2] Throughout most of the period of crystallization of the magma the ore minerals are merely one of a number of kinds of crystals and their sinking is seriously interfered with by the presence of other crystals which may tend to sink much more slowly or in rare cases perhaps even to float. In general, we must conclude that there may be a long period in the history of a crystallizing magma during which the crystals tend to sink through the liquid as a swarm, with little tendency to relative movement between the different kinds of crystals, the sinking being determined rather by the mean density of the swarm. The result of this process is, in general, not the sorting of individual minerals, but the downward movement of the heavier minerals of early crystallization as a group and the contrasting of the minerals of the basic rocks with those of the intermediate rocks and with those of the acid rocks.

[1] The lower part of the Duluth gabbro is in places 90 per cent titaniferous magnetite (Bayley, *Jour. Geol.*, II [1894], 818). Foslie recently described a Norwegian example (*Norges. Geol. Unders. Aarbok* IV [1913], 66), and many others could be cited.

[2] Assuming the densities of magnetite and orthoclase, for example, to have roughly the same relative values at high temperatures as at low temperatures, we may deduce that magnetite crystals 0.1 mm. in diameter would sink in most granitic magmas no faster than feldspar crystals 0.4 mm. in diameter.

As the more specific descriptions given later will show, sills and laccoliths of basic magma very commonly give evidence of this process when of sufficient thickness, which apparently must, as a rule, be as much as 500 to 1,000 feet or more. Much thinner sills sometimes show it[1] and much thicker sills often fail to show it, but this is nothing more than is to be expected. A small sill which is one of a large number intruded at approximately the same time into the beds of a sedimentary series will cool more slowly than a sill of equal size which is practically alone in such a series. Many thick basic sills or laccoliths are possibly not the result of instantaneous intrusion, the spreading apart of the strata being slowly accomplished so that the thickness of the sill may not have been increased much faster than the thickness of the layers of frozen magma formed on roof and floor. In such a case no differentiation of the type referred to is to be expected. Repeated violent shocks, localized escape of gases, or anything which might cause a streaming of the liquid magma may also be expected to prevent differentiation. Finally, bodies of great size in which this type of differentiation has occurred are especially liable, because of their slow crystallization, to be still partly liquid at a time of important later disturbances, and these are likely to obscure the simple density arrangement of differentiates and especially to obliterate the evidences of gradual transition of one type into another. Thus tongues of the still liquid portion may be injected into fissures and cracks formed in the crystallized portion and give the impression of having arrived as the result of a separate and distinct principal act of intrusion when really a product of differentiation practically in place.

Important effects due to the sinking of crystals are hardly to be expected in *sills* of salic magma. Such magma depends for its fluidity on its ability to retain volatile constituents, whereas the basic magmas are comparatively fluid in virtue of the nature of the silicates themselves. The intrusion of a salic magma as a sill-like body, exposing a great total surface in more or less porous rocks, favors the escape of much of its volatile material. The

[1] The very clear example of sinking of pyroxene crystals in a sill only 30 feet thick described by Iddings from Yellowstone Park must be regarded as a very exceptional case (*U.S. Geol. Survey, Mon. XXXII*, Part 2 [1889], pp. 82–84.)

resulting great viscosity may be reasonably expected to prevent important sinking of crystals.[1] Nevertheless the sinking of crystals may be of considerable importance in a very large body of salic magma which exposes to the surrounding rocks a minimum of surface in proportion to its mass.

The two processes involving relative movement of crystals and liquid—the sinking of crystals and the squeezing out of residual liquid—aid each other in a general way in the production of an arrangement of the various differentiates, such that the heaviest lies at the greatest depth, and the lighter ones at lesser depth—a gravitative adjustment.

Daly has collected from the literature and from personal study a host of instances illustrating quite convincingly the fact of gravitative adjustment.[2] The two processes mentioned above are, the writer believes, the sole instruments of its production. On a later page a discussion will be given of the working of the processes in specific cases.

The formation of zoned crystals is a process which may well be discussed at the same time. It is somewhat related to the two processes mentioned above, inasmuch as the zoning of a crystal effectively separates the earlier-formed part of the crystal from the neighboring liquid in which it formed. This fact has an important effect on crystallization and therefore on differentiation.

THE ASSOCIATION OF DIABASE AND MICROPEGMATITE

Diabases with micropegmatite interstices are very common. Sometimes the micropegmatite (granophyre) is separated as a distinct body, a granite, granodiorite, or quartz diorite in composition. This association is of fundamental importance to petrogenic theory and will be made the starting-point for a discussion of the geologic evidence supporting crystallization differentiation. It is, in many cases, clearly shown that when the diabasic (basaltic) magma was intruded as a small body and was therefore quickly

[1] In pegmatite sheets in California in which, no doubt, fluidity was maintained, Schaller has found very striking examples of the sinking of garnet crystals (personal communication from Dr. Schaller).

[2] *Op. cit.*, pp. 228 f.

chilled, it crystallized as a normal plagioclase-pyroxene diabase without quartz. On the other hand, large bodies usually show micropegmatitic interstices and often a similar salic differentiate. This contrast between the larger and the smaller bodies has led some petrologists to the opinion that the more slowly cooled, large bodies had an opportunity denied the quickly cooled bodies—the opportunity to assimilate siliceous material, whence the siliceous differentiate. Direct evidence of adequate assimilation is seldom if ever clear; its accomplishment is nearly always inferred from the existence of the acid differentiate. In the following an attempt will be made to show that the distinction often noted between the large and small bodies is solely the result of a difference in the course of crystallization dependent upon the difference in the rate of cooling. The plan will be to discuss the physical chemistry of the crystallization of investigated systems which have a direct bearing on the question and then to apply the facts and principles developed to the crystallization of basaltic magma. It may seem, perhaps, that a discussion in which physical chemistry plays so prominent a part has no place in a geological paper, but it is in reality one of the many phases of that subject which the petrologist must master if a foundation is desired on which a discussion of the crystallization of igneous magmas may be based.

THE SYSTEM DIOPSIDE-FORSTERITE-SILICA

In a recent paper the writer has published an investigation of the system diopside-forsterite-silica, in which the comparative effects of slow and of quick cooling were discussed on the basis of definite experimental results. The discussion has sufficient importance in the present connection to warrant its partial repetition here. For fuller details the original paper may be consulted.[1]

Fig. 1 is an ordinary ternary composition diagram, composition being plotted on triangular co-ordinates. Any point within the triangle represents a definite composition, definite relative proportions of the three components. The area of the triangle is divided into four *fields* by the *boundary curves*, *BFA*, etc. The field of forsterite is such that it contains all the points representing

[1] *Am. Jour. Sci.* (4), XXXVIII (1914), 207.

the composition of all the liquids which can exist in equilibrium with (are saturated with) crystals of forsterite. Similarly the field of the pyroxenes contains the composition of all the liquids which can exist in equilibrium with (are saturated with) crystals of pyroxene, the composition of the pyroxenes themselves being represented by points lying along the *conjugation line* (dotted line joining $MgSiO_3$ and diopside, between which there is an unbroken series of pyroxene mix-crystals). Any liquid whose composition is represented by a point lying on a boundary curve is in equilibrium with (saturated with) two kinds of crystals, and any liquid whose composition is represented by the point at which three boundary curves meet is saturated with three kinds of crystals. There is only one such liquid in the present system, that in equilibrium with pyroxene, tridymite, and cristobalite. Fig. 1 is therefore a *solubility diagram.*

By drawing isotherms on the figure any point can be made to represent (besides a definite composition) a definite temperature. In Fig. 1 the direction of falling temperature is merely indicated by arrows on the boundary curves, but in Fig. 2 the experimentally determined isotherms or temperature contour-lines are shown. In Fig. 2 any point within the forsterite field represents, not only the composition of a liquid saturated with forsterite, but also the temperature at which the liquid becomes just saturated with forsterite and its crystallization begins. This is also, of course, the temperature at which melting is complete, and Fig. 2 is therefore, besides a solubility diagram, a *melting-point diagram.* It is more common to refer to such diagrams as melting-point diagrams, emphasis being placed on the temperature aspect, but in complicated systems it would, for many purposes, be better to emphasize the composition aspect, solubility being the principal consideration, as will appear later in the consideration of the crystallization of magmas.

Fig. 3 is analogous to Fig. 1, showing the boundary curves somewhat displaced in order to lessen the confusion of lines and facilitate the discussion. Isotherms are omitted, but it must be considered that any point represents both a definite composition and a definite temperature.

Crystallization with perfect equilibrium.—Crystallization of a liquid of composition *M* (Fig. 3) takes place in the following manner *when perfect equilibrium prevails.* At the temperature of the point *M* the liquid becomes saturated with olivine (forsterite), and crystals of that substance begin to separate. The composition of the liquid then changes along the line *AMK* toward *K* with continued separation of forsterite. When the temperature of the point *K* is

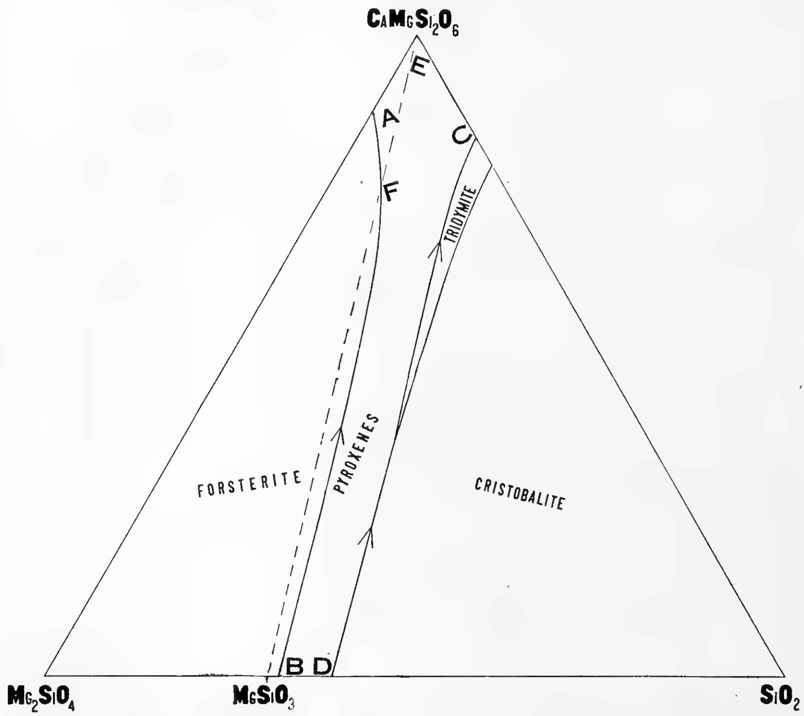

FIG. 1.—Field diagram of the system: diopside, forsterite, silica

reached, the liquid becomes saturated with pyroxene, and pyroxene of composition *L* begins to separate, i.e., pyroxene rich in magnesia. At the same time forsterite begins to redissolve and the composition of the liquid begins to change along the boundary curve from *K* toward *N*. Meantime the composition of the pyroxene separating, as well as that of the pyroxene which has already separated, changes from *L* toward *P* (becomes more calcic). When the temperature of

the point N is reached, the last of the liquid disappears, the vanishing quantity having the composition N, and the pyroxene having the composition P. The whole then consists of pyroxene of composition P and forsterite in the proportions pyroxene:forsterite= $MA:MP$.

If the composition of the original liquid had been that of the point P, the separation of forsterite would begin at the temperature

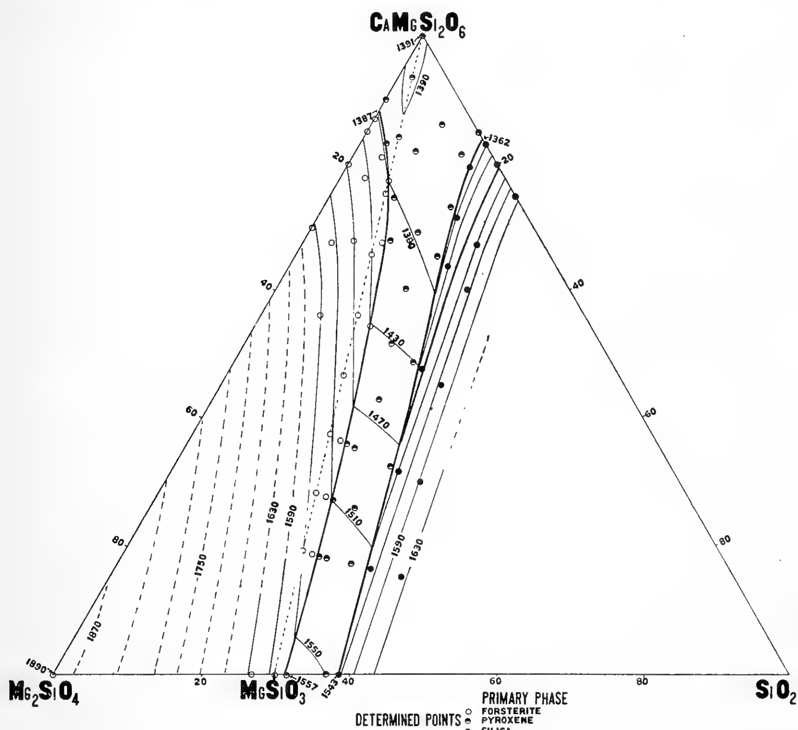

FIG. 2.—Temperature-composition diagram of the system: diopside, forsterite, silica

of the point P, and the course of crystallization would be the same as in the foregoing case except that at the temperature of the point N the last of the forsterite is redissolved and the last of the liquid is used up at the same instant and the whole consists of pyroxene of composition P. Thus all mixtures of forsterite with pyroxene P, or pyroxene P itself (i.e., any composition represented

by a point on the line AP) become completely crystalline at the temperature of the point N, the last minute quantity of liquid having the composition N. A mixture of pyroxene L with forsterite (i.e., any composition lying on the join AL) would, however, become completely crystalline at the higher temperature K, and the last minute quantity of liquid would have the composition K. A mixture of pyroxene T with forsterite (i.e., any composition lying

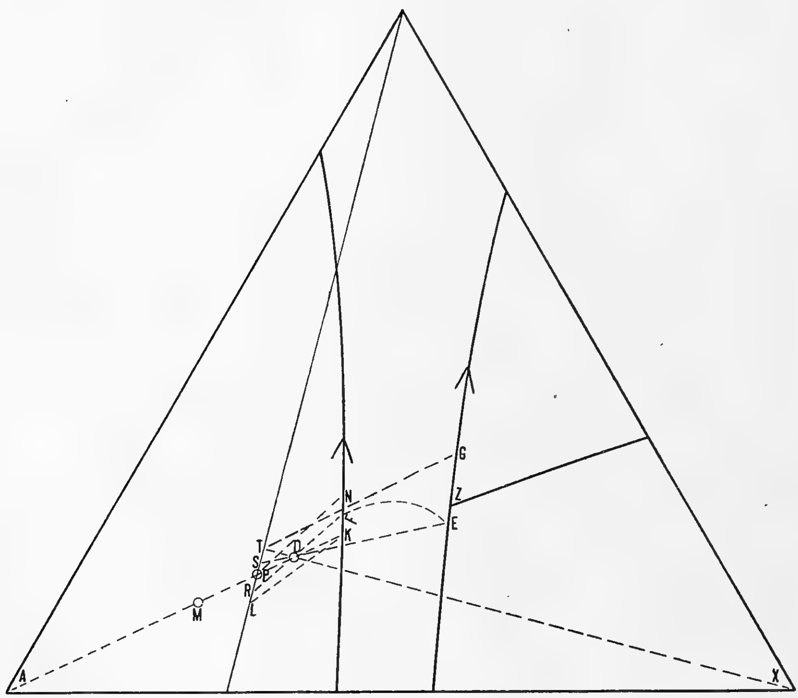

FIG. 3.—Crystallization curves for the system: diopside, forsterite, silica

on the join AT) would become completely crystalline at a temperature lower than N, and the last of the liquid would be more calcic than N.

Thus any mixture represented by a point on any line passing through A has a temperature of final crystallization different from that of a mixture lying on any other line through A, and the composition of the final liquid is different in each case. There is, moreover, in each case a continual change in the composition of the

liquid as the temperature falls, until the last drop is used up, and a concomitant change in the composition of the mix-crystals. Contrast such crystallization with the crystallization in a simple eutectic system, which is sometimes regarded as throwing light on the crystallization of igneous rocks. In the eutectic system, in all possible mixtures, the composition of the liquid continually approaches that of the eutectic mixture, and when it attains this composition there is no further change and no change in temperature until all is crystalline. The temperature of final crystallization is the same for all mixtures. The difference between these two types of systems is due principally to the existence of a mix-crystal series in the one and its absence in the other. Through consideration of eutectic systems, little aid is to be obtained toward the understanding of the crystallization of igneous rocks, in which mix-crystals are so common.

If the composition of the original liquid was that of the point D, forsterite would crystallize first as before and crystallization would follow the same course as in the two preceding cases (M and P) until the temperature of the point F is reached. At this temperature the re-solution of forsterite is complete and the mixture consists of liquid of composition F and pyroxene of composition R. The composition of the liquid now leaves the boundary curve and crosses the pyroxene field on the curve FE, pyroxene continuing to crystallize and changing in composition toward S. When the temperature of the point E is reached, cristobalite begins to crystallize. At this temperature the liquid has the composition E and pyroxene the composition S. With further lowering of temperature the composition of the liquid changes along the boundary curve from E toward Z, cristobalite and pyroxene continue to crystallize, and the pyroxene changes in composition toward T. At the temperature of the point Z, cristobalite changes to tridymite, and with further lowering of temperature tridymite and pyroxene continue to crystallize. When the temperature of the point G is reached, the liquid finally disappears; the last minute quantity has the composition G and the pyroxene has the composition T. The whole now consists of pyroxene of composition T and tridymite in the proportion pyroxene : tridymite $= DX:DT$.

Crystallization with zoning.—It has been stated that crystallization takes place after the manner described for the foregoing mixtures when perfect equilibrium is obtained. The early crystals of forsterite must be partly or completely redissolved and the crystals of pyroxene must, by diffusion in the solid, continually change composition throughout their entire mass. This result could be accomplished only with extremely slow cooling and with continuous mixing to prevent the sinking of crystals. With quicker cooling an entirely different result will be obtained.

If the cooling is at such a rate that forsterite crystals are not redissolved and pyroxene crystals once separated do not change their composition, then from *any of the liquids M, P, or D* of Fig. 3 forsterite would separate first as before until the temperature of the point K was reached, when the liquid has the composition K. At this temperature pyroxene of composition L begins to separate as before, but re-solution of forsterite does not take place in this case. The composition of the liquid therefore does not follow the boundary curve but crosses the pyroxene field and meets the boundary curve pyroxene-tridymite at a point lower than E, say G. In the meantime the composition of the pyroxene separating has changed from L to T, and the pyroxene crystals therefore show zones of continuous change of composition varying from L to T. When the temperature has reached that of the point G and the composition of the liquid is G, tridymite begins to crystallize and the composition of the liquid changes along the boundary curve pyroxene-tridymite. Meantime the composition of the pyroxene separating changes from T toward pure diopside, and final crystallization takes place only when the temperature is that of the eutectic diopside-tridymite, when the remaining infinitesimal amount of liquid has the composition of this eutectic and the crystalline phases separating are tridymite and pure diopside.

Any of the liquids M, P, or D will, then, if crystallized in this manner, consist of forsterite, tridymite, and pyroxene varying in composition from L to pure diopside. The actual amount of pyroxene approaching L in composition is relatively large; the amount approaching diopside, relatively small; the amount of pure diopside, infinitesimal. The kind of crystallization which

results in the foregoing case will be referred to in the following as *maximal zoning*.

If the cooling is extremely rapid, the mixture M may undercool to some temperature below N before crystallization begins, when it will crystallize to a mixture of forsterite and pyroxene P. It has already been shown that extremely slow cooling (with mixing) will give the same final product, no zoning of the crystals resulting in either case. There is one definite intermediate rate of cooling which will give maximal zoning. If the cooling is somewhat slower than the definite rate which gives maximal zoning, there will be opportunity for partial change in the composition of the various zones, and the mixture D, for example, instead of becoming completely crystalline only at the temperature of the eutectic diopside-silica, will be completely crystalline at a somewhat higher temperature, and the outer zone of the mix-crystals, instead of being pure diopside, will be somewhat more magnesian. If the cooling is somewhat faster than the rate which gives maximal zoning, the same result will be accomplished, this time because of slight undercooling maintained throughout crystallization. Any intermediate rate of cooling differing from the rate which gives maximal zoning will, therefore, give some zoning, but with a smaller range of composition between the inner and outer zones, this being accompanied by a smaller range of temperature of crystallization and a smaller range in the progressive change in the composition of the liquid than that shown in maximal zoning. It is obvious that the rate of cooling has a fundamental effect on the course of crystallization in the mixtures of the system investigated and indeed must have a similar effect on the crystallization of any mixtures from which mix-crystals separate.

Crystallization when the crystals are free to sink in the liquid.—Even when the cooling is extremely slow, approaching or even realizing the rate required to produce perfect equilibrium, the sinking of crystals is a factor entering into the problem and tending to produce the same general result as the intermediate rates of cooling which give rise to zoning. The continual downward movement of growing crystals accomplishes their removal from that part of the liquid from which they crystallized. The effect

on the liquid is similar to that produced by zoning, for zoning accomplishes a cutting off of an early crystal from the liquid by the formation of a zone of different composition about it.

A mass of liquid of composition M is cooled uniformly in all parts and forsterite crystals are imagined to sink from the upper to the lower layers. When the temperature K is reached, the liquid has the composition K in all parts, but it holds only a few olivine (forsterite) crystals in its upper layers and many in its lower layers.[1] The total composition of a certain upper layer could, then, be represented by the point D, and of a certain lower layer by some point between M and A. Imagine, for simplicity, that further sinking of crystals is prevented; then, as the temperature falls below K the separation of pyroxene and the re-solution of forsterite proceed in all layers. When the temperature F is reached, the solution of forsterite is complete in the upper layer of composition D, but is still incomplete in the lower layer. With further cooling, re-solution of forsterite continues in the lower layer until at the temperature N all the liquid is used up. *The lower layer is now completely crystalline*, consisting of a mixture of pyroxene P and forsterite, *but the upper layer is still partly liquid*, for, in the meantime, the composition of liquid in this layer has crossed the pyroxene field on the curve FE and the liquid is not entirely used up until the temperature G, much lower than N, is reached. When the temperature G is reached, the upper layer then consists of pyroxene of composition T and tridymite. There is, therefore, not merely a marked difference in composition between the lower and the upper parts of the crystalline product, but also *a partly different assemblage of minerals.*

It is important to note that as the temperature falls below that of the point F the composition of the liquid portion, which up to this point has been uniform from top to bottom, now pursues a different course in the upper layers from that followed in the lower layers. At a temperature a few degrees below F the composition of the liquid in the lower layers is represented by a point between F

[1] This sinking of olivine crystals is not an imaginary process but has been obtained in crucibles containing these melts (N. L. Bowen, *Am. Jour. Sci.* [4], XXXIX [1915], 175).

and N and that in the upper layers by a point on the curve FE. A squeezing out of the liquid from such a mass might bring these different liquids together and draw them out into streaks which though perfectly miscible would have only a limited opportunity to mix before crystallizing, and would thus give a result simulating primary banding, a rather rare but quite definite phenomenon of igneous rocks.

It has been supposed in the foregoing that only crystals of forsterite settle out, but it may readily be imagined that in a certain limited portion nearly all the forsterite has settled out at the time when the crystallization of pyroxene begins and that there is opportunity for the sinking of pyroxene crystals. Even in layers having an abundance of forsterite crystals, pyroxenes will crystallize and join the sinking swarm. The result is an enrichment of the lower layers in the more magnesian pyroxenes of early separation with a consequent *raising* of their temperature of final consolidation somewhat above N, whereas the liquid in the upper portion is continually enriched in more calcic pyroxene with a consequent *lowering* of its temperature of final consolidation below G. The composition of the liquid is, in fact, continually *offset* toward the eutectic diopside-silica. In an ideal case the composition of the last minute quantity of liquid would reach the eutectic diopside-silica, but in any actual case the eutectic will not be attained, the last of the liquid being used up before the temperature and composition of the eutectic is reached. The temperature of final consolidation of the residual liquid and its composition will lie at some point intermediate between G and the eutectic diopside-silica, approaching closer to the latter according as the opportunity for the sinking of crystals increases. The opportunity for the sinking of crystals is directly related to the rate of cooling. It is obvious, therefore, that the rate of cooling is of fundamental significance in controlling the ultimate products of crystallization.

In a system with simple eutectic relations, settling of crystals could make no difference in the temperature of final consolidation of the last of the liquid, nor in its composition.

Crystallization with straining off of the liquid.—A process involving the separation of crystals and liquid by the squeezing out of

residual liquid will evidently work in the same direction as the two processes of zoning and sinking of crystals, but it differs in not having the same continuous nature, being rather an incident in the history of the magma, though, no doubt, repeated several times in some cases.

THE SYSTEM ANORTHITE-FORSTERITE-SILICA

After discussing the crystallization of liquids of the foregoing system involving an olivine, a series of pyroxenes and silica, a system involving a feldspar, an olivine, a single pyroxene, and silica will now be considered. This is the system anorthite-forsterite-silica recently investigated by Andersen.[1]

The equilibrium diagram is shown in Fig. 4.

Crystallization of liquid of composition x, a mixture of clino-enstatite and anorthite, may take place in any of the following ways. With perfect equilibrium and sinking of crystals prevented, forsterite separates first and the composition of the liquid changes along the line xy. At the temperature y the pyroxene clino-enstatite begins to separate and forsterite to redissolve, the liquid changing in composition along the boundary curve FM. When the temperature M is reached, the last of the liquid (composition M) and the last of the forsterite disappear and the whole consists of the pyroxene, clino-enstatite, and anorthite. The same final result would be obtained if the liquid were cooled very rapidly, that is, if considerable undercooling occurred.

If, however, forsterite crystals settled from the upper to the lower layers, the consequence would be that at the temperature M the liquid would be used up in the lower layers before all the forsterite was redissolved and the whole lower portion would be crystalline, consisting of the *olivine, forsterite, the pyroxene, clino-enstatite, and anorthite.* The earlier crystals of olivine might collect in a separate layer, peridotite-like in nature. In the upper layers, however, the little forsterite remaining is redissolved while some liquid of composition M still remains and, after complete solution of forsterite, the liquid changes in composition along MN, clino-enstatite and anorthite separating until at the temperature N

[1] Olaf Andersen, *Am. Jour. Sci.* (4), XXXIX (1915), 407.

tridymite also separates and the whole crystallizes to a mixture of *the pyroxene, clino-enstatite, anorthite, and silica.* Thus the sinking of crystals has brought about a partly different assemblage of minerals in the upper and lower layers, a difference in temperature of final consolidation and in the composition of the final liquid of these layers. It is evident that even if the cooling were too

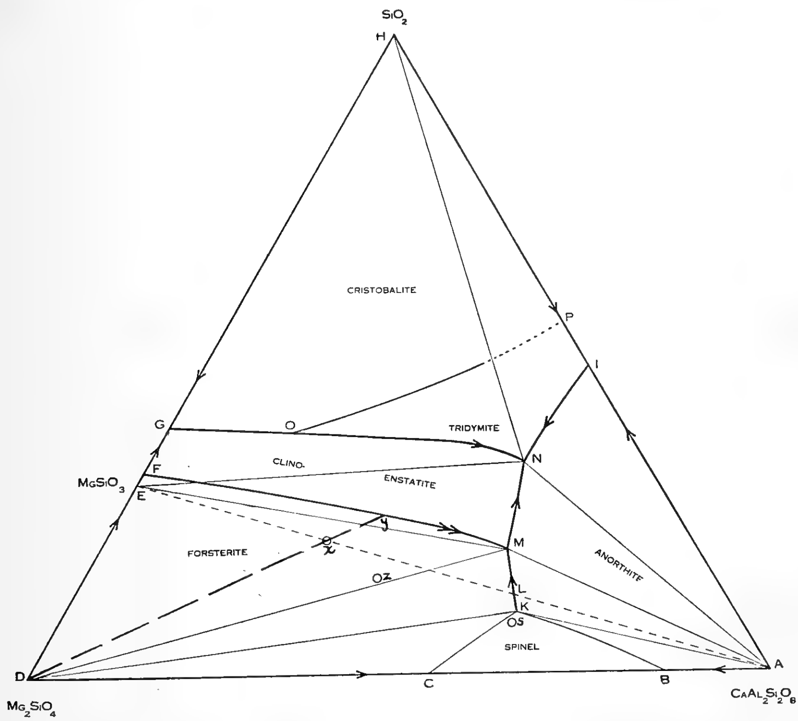

FIG. 4.—Equilibrium diagram of the system: anorthite, forsterite, silica, after Andersen, *Am. Jour. Sci.* (4), XXXIX (1915), 437.

quick to permit of any considerable sinking of crystals, the same consequence to the composition of the final liquid would result from the formation of a reaction rim of clino-enstatite about the forsterite crystals and their protection thereby from further resorption by the liquid.

If the composition of the original liquid were that of the point z, crystallization of the perfect equilibrium type and also crystallization

as a result of very rapid cooling would give a mixture of forsterite, clino-enstatite, and anorthite. Sinking of crystals would have the same result as in the case of the liquid x, but the total amount of forsterite formed would be greater and a considerable peridotite-like layer might be formed. If the composition of the liquid were that of the point s, slow crystallization with perfect equilibrium and no sinking of crystals and also very rapid cooling would give a final product of forsterite, clino-enstatite, and anorthite rather rich in forsterite but entirely free from spinel. Slow crystallization with sinking of crystals would remove the early-formed spinel crystals from that part of the liquid in which they formed and thereby prevent the re-solution which takes place with perfect equilibrium. These spinel crystals, collecting with the earlier forsterite crystals, would give a mass simulating a *spinel-bearing peridotite.*

It is to be noted that during the crystallization of the liquid x in which complete re-solution of olivine takes place with perfect equilibrium, there is a stage at which olivine crystals constitute about *15 per cent of the whole mass.* In case forsterite crystals sank during the crystallization of this mixture, the liquid would be of the same composition in all parts until the temperature y was reached, after which the liquid in the part containing olivine crystals would change in composition in a different manner from the parts of the liquid free from olivine crystals. The latter would depart from the boundary curve FM, cross the clino-enstatite field, and become completely crystalline when the temperature N was reached. The former would follow the boundary curve and become completely crystalline at the higher temperature M. During the period when the liquid in contiguous parts was forced to follow different courses by the presence of olivine crystals in the one part and their absence in the other, it seems likely that a squeezing out of the liquid might give rise to phenomena similar to those noted in the rather rare natural case of the intrusion of a heterogeneous magma—might, in fact, give a product simulating a banded gabbro. The liquids concerned would not, however, be immiscible.

It may be pointed out also that certain layers into which considerable olivine had settled would be peridotitic (lherzolitic) in

total composition (liquid and crystals) and at the same time would for a considerable period contain nearly or quite 50 per cent liquid. With such a proportion of liquid it would be eruptible as a whole and might be injected as dykes of peridotite into adjacent rocks, if we imagine these simplified magmas occurring in the crust of the earth. The eruptibility of this mixture would be increased by the fact that the injection must take place from a position of higher to one of lower pressure and, as a result of the lowering of pressure, re-solution of some crystals would take place. It appears likely, however, that this re-solution must be considered on the whole of relatively small importance and merely as aiding somewhat in the eruption of parts of a magma enriched in sunken crystals. We know little about the volume-changes and heat-effects involved in the solution of silicates in silicate liquids, but such indications as are to be gained from the same factors for the melting of single silicates suggest that the effect of relief of pressure in bringing about re-solution would be moderate. The great importance of this process assumed by Schweig[1] is unlikely. It appears also to be unnecessary to make this assumption in order to explain the facts for which it was proposed.

An increase in the eruptibility of a part enriched in sunken crystals through re-solution of the crystals in the hotter liquid appears to be also of limited importance. The factors limiting this action will be discussed on a later page.

CRYSTALLIZATION IN SYSTEMS INVOLVING THE PLAGIOCLASES

In natural rocks the plagioclases are a very important mix-crystal series. We shall now consider a system related to both those already discussed, involving the single pyroxene diopside, and the plagioclase mix-crystal series instead of the single plagioclase anorthite.[2]

CRYSTALLIZATION OF THE PLAGIOCLASES IN THE BINARY SYSTEM

Experimental work has confirmed the opinion that the plagioclases belong to Type I of Roozeboom's classification of mix-crystal

[1] *Neues Jahrb.*, Beil. Band XVII (1903), 563.

[2] N. L. Bowen, "The Crystallization of Haplobasaltic, Haplodioritic and Related Magmas," *Am. Jour. Sci.* (4), XL (1915), 161.

series.[1] It has also been possible to determine accurately the position of both solidus and liquidus so that the exact composition of solid and liquid in equilibrium with each other in the binary system is known[2] (see Fig. 5). The results, given in the paper referred to, show that the liquids are always very much richer in albite than

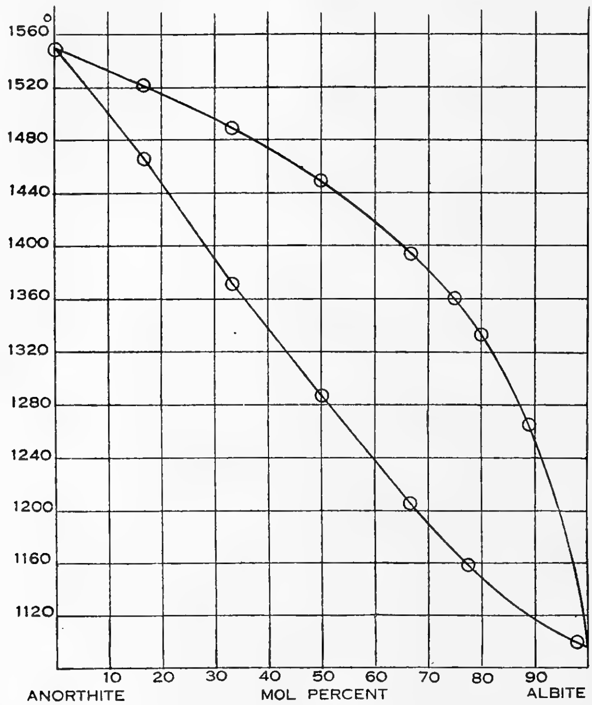

Fig. 5

the crystals with which they are in equilibrium. In discussing the crystallization of a liquid the consequences of zoning or the sinking of crystals were pointed out.[3] On account of the great viscosity of their liquids and consequent ease of undercooling, the artificial plagioclases prepared did not exhibit zoning, being different in this

[1] Day and Allen, "The Isomorphism and Thermal Properties of the Feldspars," *Carnegie Institution of Washington, Publ. No. 31.*

[2] N. L. Bowen, "The Melting Phenomena of the Plagioclase Feldspars," *Am. Jour. Sci.* (4), XXXV (1913), 583.

[3] *Ibid.*, p. 597.

respect from the pyroxenes described. Nevertheless the exact determinations of solidus and liquidus affords a sound foundation for the discussion of zoning in the plagioclases.

When the cooling is at such a rate that zoning results, the composition of the liquid is continually offset toward pure albite. There is a certain definite rate of cooling which gives maximal zoning, in which case the last minute quantity of liquid as well as the outer zone of the crystal has the composition of pure albite, whatever the total composition of the original material may have been. With a somewhat quicker rate of cooling, the range of zoning is not so great owing to a moderate degree of undercooling, and when the undercooling is very great there is no zoning at all. With a rate of cooling somewhat slower than that which gives maximal zoning, the range of zoning is, again, not so great, owing to the opportunity afforded for partial adjustment of composition between the various zones, and when the cooling is extremely slow there is no zoning at all on account of perfect adjustment. Since zoning is absent in this latter case of extremely slow cooling, there is no continuous offsetting in the composition of the liquid in the direction of pure albite from that cause, but the great opportunity for the settling of crystals afforded by slow cooling enters into the problem with the tendency to produce the same result. It is therefore clear that the tendency toward the continuous offsetting in the composition of the liquid increases with slow cooling.

CRYSTALLIZATION IN THE SYSTEM DIOPSIDE-ANORTHITE-ALBITE

It was with the purpose of showing the manner in which the properties of plagioclase mixtures are carried into polycomponent systems that the investigation of the ternary system diopside-plagioclase was undertaken, and fortunately definite results were obtained.

Fig. 6 is the equilibrium diagram. All liquids whose composition is represented by points in the field *BCED* are capable of existing in equilibrium with plagioclase, and all liquids in the field *AED* may exist in equilibrium with diopside; consequently all liquids represented by points lying along the boundary curve *DE* can exist in equilibrium with both plagioclase and diopside. In

Fig. 7 the temperature relations of the fusion surface are indicated by means of isotherms or temperature contours. The isotherms indicate the shape of the fusion surface in the same manner that elevation contours represent the shape of the surface of the ground on a topographic map, the contour interval being a certain number of degrees instead of a certain number of feet. Any point on Fig. 7

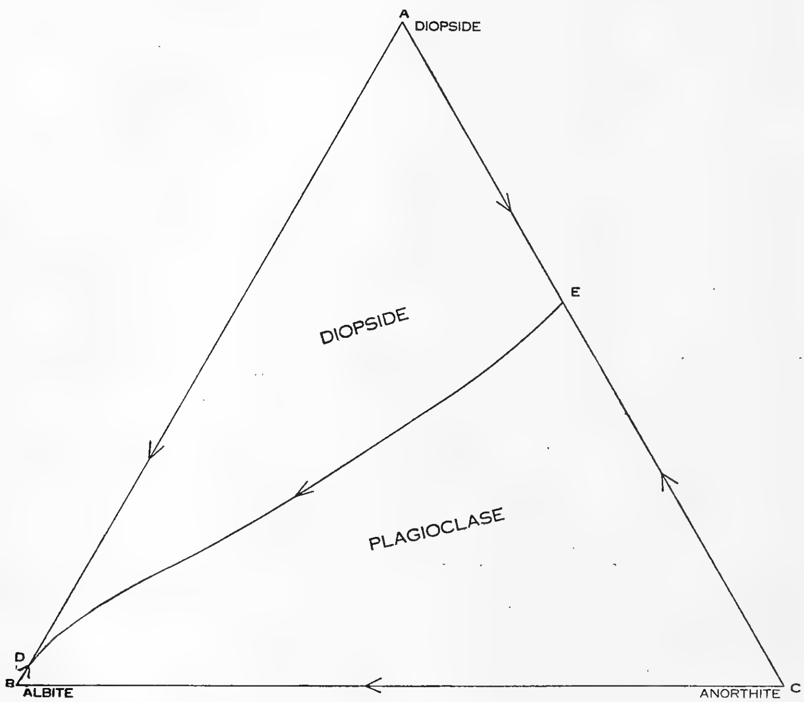

Fig. 6.—Field diagram of the system: diopside, plagioclase

represents, therefore, not only a definite composition, but also a definite temperature. Any point within the plagioclase field represents the composition of a liquid which can exist in equilibrium with plagioclase and at the same time represents the temperature at which that liquid becomes just saturated with plagioclase.

The crystallization of typical mixtures may now be considered. On account of the occurrence of a mix-crystal series it is necessary to distinguish, as was done in the system involving the pyroxene

series, between the two types of crystallization, that in which perfect equilibrium is obtained and that in which zoning occurs.[1]

Crystallization with perfect equilibrium.—A liquid of composition *F* (Fig. 8) which contains 50 per cent labradorite (Ab_1An_1) and 50 per cent diopside begins to crystallize at 1275°, diopside separating first. As the temperature falls the composition of the liquid

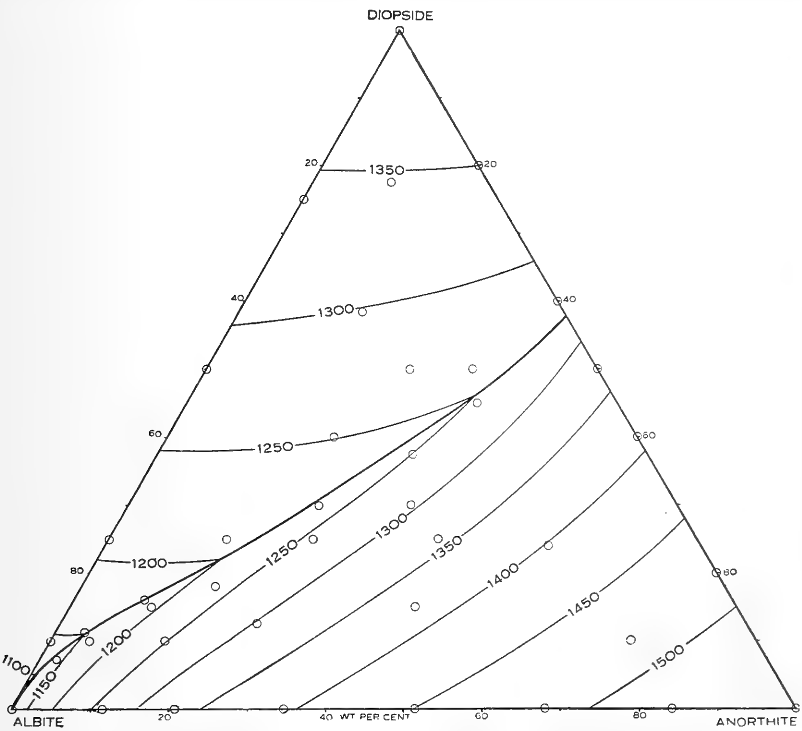

FIG. 7.—Equilibrium diagram of the system: diopside, plagioclase

changes from *F* along *AFG* (directly away from diopside), and when the temperature 1235° is reached the mass consists of 17 per cent diopside crystals and 83 per cent liquid of composition *G*. At this temperature bytownite of composition *H* (approximately Ab_1An_4) begins to crystallize and the composition of the liquid changes along the boundary curve *DE* toward *D*. Both diopside

[1] These correspond with "Erstarrung zweiter Art" and "Erstarrung erster Art" respectively of Schreinemakers (*Zs. phys. Chemie*, L [1905], 189–90).

and plagioclase continue to separate and the plagioclase crystals, not only those separating at any instant, but also those which had formerly separated, continually change in composition—become richer in albite. At 1220° the whole mass is made up of 37 per cent diopside crystals, 25 per cent labradorite crystals of composition *L* (Ab_1An_2), and 38 per cent liquid of composition *K*. As the temperature falls still lower the liquid gradually decreases in amount

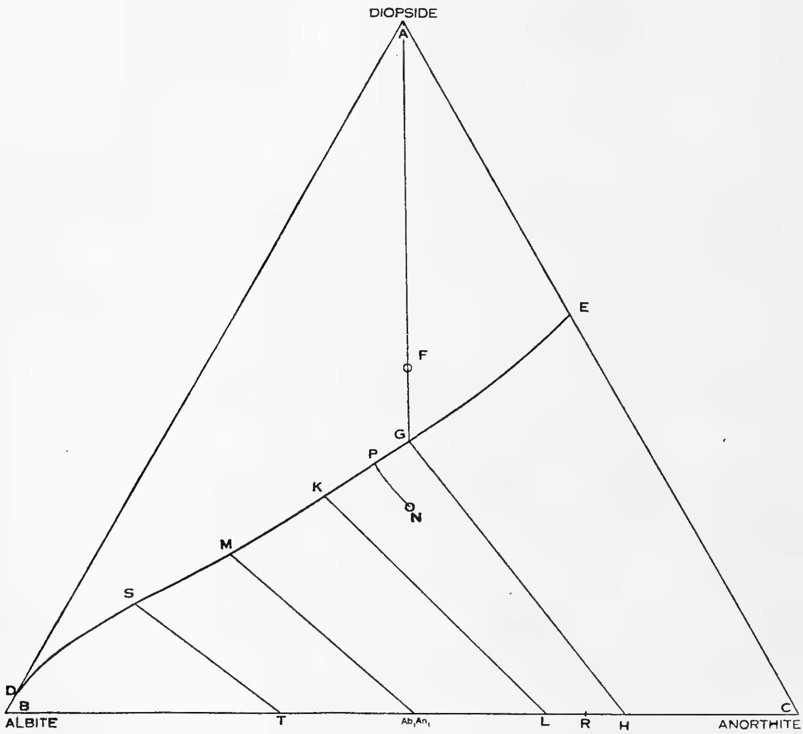

FIG. 8.—Crystallization curves for the system: diopside, plagioclase

and continually changes in composition until at 1200° it is all used up, the last minute quantity having the composition *M*. In the meantime diopside and plagioclase crystals have been separating and the plagioclase has been changing continuously in composition until at 1200°, when the last of the liquid disappears, the composition of the feldspar is Ab_1An_1. The whole mass now consists of 50 per cent diopside and 50 per cent Ab_1An_1.

If the composition of the original liquid is that of the point *N*, which consists of 70 per cent plagioclase Ab_1An_1 and 30 per cent diopside, crystallization begins with the separation of bytownite (Ab_1An_4 approximately) at 1302°. As the temperature falls the composition of the liquid changes, along the curve *NP*, plagioclase continues to separate, and its composition changes until at 1230°, when the liquid has the composition *P*, the feldspar has the composition $R(Ab_1An_3)$. With further cooling, the composition of the liquid changes along the boundary curve, diopside separating as well, and, as in the preceding case, the last of the liquid disappears at 1200°, when it has the composition *M*, and the plagioclase crystals have changed gradually to Ab_1An_1. The whole mass now consists of 70 per cent plagioclase Ab_1An_1 and 30 per cent diopside.

The two mixtures *F* and *N* are merely random mixtures of Ab_1An_1 and diopside on opposite sides of the boundary curve. All mixtures of Ab_1An_1 and diopside, therefore, become completely crystallized at 1200°, when perfect equilibrium is obtained, and the last minute quantity of liquid has the composition *M*, which, expressed mineralogically, is 78 per cent Ab_4An_1 and 22 per cent diopside.

All mixtures of diopside with Ab_1An_2 become completely crystallized at 1220°, the last of the liquid having the composition *K*. Mixtures of Ab_2An_1 and diopside are completely crystalline at 1178° and the last of the liquid has the composition *S*. Thus a mixture of diopside with any plagioclase has a temperature of final consolidation different from that of a mixture of diopside with any other plagioclase, and to speak of a eutectic between diopside and plagioclase is to use a meaningless phrase.[1] "The eutectic of plagioclase and pyroxene" is even more emphatically meaningless when the pyroxene referred to is a complex mix-crystal such as augite.

Crystallization with great undercooling.—The same final product will be obtained in each of the foregoing cases if the liquid is greatly

[1] Even the definite temperature of final consolidation of mixtures containing a certain definite plagioclase is not properly to be called a eutectic. Thus 1200° is not the eutectic temperature of Ab_1An_1 and diopside, and the liquid *M* is obviously not the eutectic mixture of Ab_1An_1 and diopside.

undercooled. Thus if the liquid F is cooled very quickly to a temperature below 1200°, crystallization, if it takes place at all, will give 50 per cent diopside and 50 per cent plagioclase of the uniform composition Ab_1An_1 (unzoned).

Crystallization with zoning.—When the cooling is too rapid to give crystallization of the perfect equilibrium type and yet not rapid enough to give the great degree of undercooling referred to in the foregoing, the formation of zoned crystals of plagioclase will result. According as the one or the other of the above-named rates of cooling is approached the degree of zoning is reduced to a minimum. With a certain intermediate rate of cooling maximal zoning results. In this case a crystal once separated suffers thereafter no change of composition, the liquid disregarding crystals which have already formed, so that the crystallization of the liquid may be regarded as beginning anew at each instant.

The effect of this action may be realized by considering that during the crystallization of the liquid F, as already outlined, the liquid portion is separated from the crystalline portion at a temperature of, say, 1220°. At this temperature the liquid has the composition K and we shall imagine that this separated liquid is crystallized under perfect equilibrium conditions. Instead of becoming completely crystalline at 1200°, as it would if the crystals had not been removed, it now becomes completely crystalline only at 1178°, and the final liquid, instead of the composition M, has the composition S, i.e., is much richer in albite. If the virtual separation of liquid from crystals is a continuous process accomplished through the intervention of zoning, it is plain that the offsetting in the composition of the final liquid is limited only by the eutectic albite-diopside which it actually attains in the case of maximal zoning. This fact is true, not only of the special liquids to which reference has been made, but of *any mixture of anorthite, albite, and diopside whatsoever.*

Crystallization with subsidence of crystals.—The sinking of crystals of plagioclase in a mass of liquid which is very slowly cooled will obviously affect the upper layers from which the crystals have settled in the same manner that zoning affects the residual liquid. The upper layer is continually enriched in albite and its tempera-

ture of final consolidation continually lowered. This latter effect is impossible in any eutectic system. Moreover, the temperature of final consolidation of the lower layers is continually raised.

The effect of rate of cooling summarized.—When the liquid is very quickly cooled, it crystallizes quickly, if at all, and with little or no tendency to an offsetting in the composition of the liquid. If it is cooled moderately slowly, zoning of the plagioclase causes a continual enrichment of the residual (interstitial) liquid in albite. If it is cooled still more slowly, sinking of plagioclase causes a similar continual enrichment of the residual (upper) liquid in albite. In favorable cases the final liquid may be *more than 90 per cent albite* even although the original mixture were, say, 50 per cent diopside and 50 per cent bytownite.

CONCLUSIONS TO BE DRAWN FROM THE INVESTIGATED TERNARY SYSTEMS

If we now combine the information furnished by the investigated systems, important conclusions may be drawn with regard to the crystallization of basaltic magma under various conditions. Instead of the simple pyroxene diopside, present in the mixtures of the last system discussed, we may consider one of the intermediate pyroxenes, which melt with decomposition and the formation of olivine, to be present in addition to plagioclase. Rapid cooling of such a liquid would give merely plagioclase and pyroxene. On the other hand, slow cooling permits radical variation from this simple result. The early formation of olivine brings about an excess of free silica in the residual liquid if any process intervenes to prevent the resorption of the olivine by the liquid. The early formation of very calcic plagioclase brings about an enrichment of the liquid in albite if anything intervenes to prevent the continual alteration in the composition of the crystals by interchange with the liquid. Finally the early formation of magnesia-rich pyroxene brings about an enrichment of the liquid in diopsidic pyroxene if similar conditions intervene.

The sinking of crystals affords a means of continually separating crystals from the part of the liquid in which they formed and is therefore a process which will give the results just outlined. If, therefore, the mixture of plagioclase and pyroxene referred to were cooled

slowly and continual sinking of crystals occurred, the inevitable result would be a body consisting of calcic plagioclase, olivine, and magnesian pyroxene in its lower parts (i.e., of a gabbroidal nature) and of sodic plagioclase approaching albite, diopsidic pyroxene, and free silica in its upper parts (i.e., of a granitic nature), with various intermediate types in the intermediate layers. If the freedom of sinking of crystals were somewhat restricted, one of these intermediate types, say a granodiorite or a diorite, would occur as the uppermost differentiate, the limit of the process under these less favorable conditions. The composition of the residual liquid might, moreover, have been similarly affected by zoning of crystals even if there were no opportunity for the sinking of crystals, and in this case the *interstitial* material of late crystallization would be the same salic material as that found in the upper layers when sinking of crystals took place. If a certain amount of both zoning and sinking of crystals took place, a body would result showing the salic differentiate both as interstitial material and as a separate upper layer. The same possibilities of the formation of peridotite and spinel-bearing peridotite obtain in this case with the same conditions favoring their formation as those discussed under the anorthite, forsterite, silica system.

It has been possible, then, to deduce from facts ascertained experimentally the crystallization with quick and slow cooling of mixtures which give results closely analogous to the occurrence observed in nature of diabase in small dykes and small sills (quickly cooled) and of diabase with micropegmatite interstices or a granitic or granodioritic differentiate in larger bodies (slowly cooled). There are many differences and complications in the natural magma in the matter of details, but it is clear that the broad scheme is well understood and that crystallization is the sole control. There is no necessity for assuming that assimilation of siliceous material is essential to the formation of the salic differentiate, nor that its separation is accomplished by the process of liquid immiscibility. The Palisade diabase sill with its "ledge" rich in sunken olivine crystals near the base and the micropegmatite interstices in the main mass of the diabase is a case in which the working of the process is clearly exhibited.[1]

[1] J. V. Lewis, *Ann. Rept. State Geologist, New Jersey*, 1907, p. 125.

THE IMPORTANCE OF THE FORMATION OF BIOTITE

Considerable importance has been attached in the foregoing to the early separation of olivine, but there are many cases of diabases with micropegmatite interstices, or even a separate salic differentiate, where there is good evidence that no early separation of olivine took place. Whence, then, came the free silica (quartz)? It is, of course, possible in some instances that the magma as intruded was a quartz-diabase magma; yet in many cases the quickly cooled, small dykes show no quartz, and prove clearly that the magma was of normal, basaltic type. There must be some process which results in the separation of free silica as quartz entirely independent of the separation of olivine in some cases, supplementing it in others. This process is that involved in the separation of the mica, biotite. The femic material which separates as pyroxene from the more basic diabase, and dominantly as hornblende in rocks of intermediate composition, appears in solid solution with alkalic molecules as biotite-mica in the salic differentiates. It is frequently noted, moreover, that when the salic material occurs rather as interstitial micropegmatite, the micropegmatite liquid where it bordered against pyroxene has reacted with it to form biotite. This difference between the natural magma and the artificial melts in which a diopsidic granite was the salic differentiate is plainly the outcome of the fact that the artificial melts are anhydrous whereas the magma contains various volatile constituents. The formation of hornblende and still more of the micas, with their essential content of water and often of fluorine, is the result of an increasing concentration of volatile constituents.

The change in the nature of the ferromagnesian constituent with the progress of crystallization has not been investigated experimentally on account of the great difficulty of dealing with iron-bearing minerals. The principles involved are, however, well illustrated in certain investigated systems and will be pointed out before proceeding to a discussion of the processes taking place.

EQUILIBRIUM IN SILICATE LIQUIDS

It is sometimes possible to obtain, from a study of the crystallization of certain mixtures, definite evidence of equilibrium reactions in the liquid.

If a liquid of composition $MgSiO_3$ is cooled, the first compound which separates is Mg_2SiO_4 (forsterite). There is, therefore, direct evidence that the compound $MgSiO_3$ is partly dissociated in the liquid state in such a manner that Mg_2SiO_4 is one of the dissociation products. A possible equation for the equilibrium reaction might be written $2MgSiO_3 \leftrightharpoons MgSiO_4 + SiO_2$. There may be, indeed probably there is, further dissociation according to the reaction $Mg_2SiO_4 \leftrightharpoons 2MgO + SiO_2$, but of this there is no direct evidence. There is certain evidence of the presence of Mg_2SiO_4 because that compound first crystallizes from the liquid; i.e., of all the possible compounds which might exist in the liquid, Mg_2SiO_4 first exceeds its solubility. The actual amount of the compound Mg_2SiO_4 may be quite small compared with the amount of undissociated $MgSiO_3$, but the solubility of the latter very much greater. Indeed, it may be stated that, on account of the higher melting-point of Mg_2SiO_4, it may be expected to exceed its solubility before $MgSiO_3$, even if present in the liquid in smaller amount.

If a mixture of Mg_2SiO_4 and anorthite, containing approximately twice as much anorthite as forsterite, is melted and then cooled, the mineral which crystallizes first is neither anorthite nor forsterite, but spinel, $MgAl_2O_4$.[1] There is, therefore, evidence of an equilibrium reaction in the liquid of which one of the products is $MgAl_2O_4$. It is uncertain what other products may be present, but of all those present the compound $MgAl_2O_4$ exceeds its saturation limit first. Magnesia was added to the mixture in the form Mg_2SiO_4 and separates from it as $MgAl_2O_4$, but this should not be construed as meaning that magnesia has a *greater affinity* for alumina than for silica. The actual amount of unchanged Mg_2SiO_4 in the liquid may have been many times as great as the amount of $MgAl_2O_4$,[2] but the solubility of Mg_2SiO_4 correspondingly greater. Solubility is the controlling factor. The solid phases (crystals) which separate from any liquid mixture of silicates cannot be regarded, therefore, as giving unqualified evidence of the relative

[1] Olaf Andersen, *op. cit.*, p. 436.

[2] The separation of $MgAl_2O_4$ even when these molecules occur in the liquid in relatively small amount is connected with its high melting-point.

affinity of the various oxides for one another.[1] Some molecules not represented in the crystalline product, because relatively very soluble, may have been present in the liquid in much greater concentration than those molecules which happened to separate as crystals. Only a complete knowledge of the relative proportions of all the compounds existing *in the liquid* could decide the question of relative affinities. Thus it is sometimes stated that potash has a greater affinity for silica than soda because certain rocks contain orthoclase ($KAlSi_3O_8$) and nephelite (principally $NaAlSiO_4$), but all that can be safely deduced from the association of these two minerals is that among all the compounds in the liquid magma these two were relatively insoluble under the conditions there prevailing. There may have been much more combination in the liquid into the molecules $NaAlSi_3O_8$ and $KAlSiO_4$. That there was at least some combination of this kind is plainly brought out by the fact that the orthoclase is never free from albite, nor the nephelite from kaliophilite.

For a similar reason statements concerning the relative strength of acids, e.g., silicic and carbonic acids, should be guarded. At first sight the reaction $BaCl_2+H_2SO_4=BaSO_4+2HCl$ might be considered to indicate that sulphuric acid is a stronger acid than hydrochloric acid, but we may immediately write the reaction $Ag_2SO_4+2HCl=2AgCl+H_2SO_4$, which, on the same basis, would prove the opposite. As a matter of fact, it is the extremely limited *solubility* of barium sulphate and silver chloride which determines the fact that both these reactions proceed practically to completion. If silicic and carbonic acids occur together in a solution, all we know is that both will exist in a very great variety of combinations; those with which we shall become acquainted in the crystalline state will be those which happen to exceed their saturation limit under the particular conditions.

It is clear then that in the discussion of the crystallization of a magma references to relative affinities should, in general, be avoided for the reason that the crystalline (least soluble) phases give no evidence on that subject. The crystalline product may,

[1] See Johnston and Niggli, "The General Principles Underlying Metamorphic Processes," *Jour. Geol.*, XXI (1913), 506–8, 513.

however, give evidence of the existence of certain equilibrium reactions in the liquid, though leaving in doubt the relative proportions of the various compounds taking part in the equilibrium.

EQUILIBRIA IN MAGMAS INDICATED BY MINERALS PRECIPITATED

We may now return to the commonly observed association of diabase and its salic differentiate. The principal minerals of the salic differentiate are orthoclase, sodic plagioclase, quartz, and biotite. It appears that special attention has seldom been called to the fact that in the *most siliceous rocks* (granites, granodiorites, etc.) considerable proportions of the alkalis and of the ferromagnesian constituents are represented, in the mineral biotite, in their *least siliceous* combinations.[1] A consideration of the equilibria in the magma indicated by this mineral grouping affords an explanation of the formation of such a differentiate when diabase magma is slowly cooled, and at the same time leads naturally to a conception of one of the modes of origin of the most important group of the alkaline rocks, the nephelite syenites and their derivatives.

Imagine a large body of basaltic magma slowly cooled and crystallizing in such a manner that the early-formed pyroxenes and more calcic plagioclases sink slowly in the liquid. The result is a continual enrichment of the part still liquid in alkaline feldspars. There is also a continual enrichment in the volatile constituents such as water, CO_2, S, Cl, etc. From this liquid are precipitated the minerals of the salic differentiate including alkaline feldspars, quartz, and biotite. The chemical characters of these minerals give direct evidence of a number of equilibrium reactions in the liquid. The most important of these reactions involve the breakdown of part of the polysilicate molecules as follows:

$$KAlSi_3O_8 \leftrightarrows KAlSiO_4 + 2\ SiO_2$$
$$NaAlSi_3O_8 \leftrightarrows NaAlSiO_4 + 2\ SiO_2$$

There must also exist such equilibria as the following:

$$NaAlSiO_4 + H_2O \leftrightarrows HAlSiO_4 + NaOH$$
$$KAlSiO_4 + H_2O \leftrightarrows HAlSiO_4 + KOH$$

[1] See Iddings, *Igneous Rocks*, I, 133.

and, doubtless,

$$NaOH + HCl \leftrightharpoons NaCl + H_2O$$
$$2\,NaOH + H_2S \leftrightharpoons Na_2S + 2\,H_2O$$
$$2\,NaOH + CO_2 \leftrightharpoons Na_2CO_3 + H_2O$$

with similar reactions for the corresponding potash compounds, besides very complicated equilibria between the molecules S, SO_2, SO_3, C, CO, CO_2, H_2S, H, H_2O, O, HCl, Cl, etc.

Though we understand very little about the exact form in which iron and magnesia enter into the micas, it appears that there is the same tendency to partial breakdown from the more siliceous metasilicate to the less siliceous orthosilicate.

There is no doubt that the increased concentration of water in the magma at this stage exerts a strong influence in promoting this breakdown of the polysilicate molecules of the alkalis and the metasilicates of iron and magnesia into the orthosilicate molecules with setting free of silica, an action which may be compared with hydrolysis. Niggli[1] ascribes such action to water in discussing the rocks of Electric Peak and Sepulchre Mountains described by Iddings.[2] In some of the deep-seated rocks of Electric Peak quartz and biotite occur, whereas they are absent in surface rocks of the same composition at Sepulchre Mountains. This is ascribed to loss of water by the surface rocks. Niggli states: "Zugleich hat dieses Wasser offenbar einen Teil der Kieselsäure aus den Verbindungen gedrängt und so in der Tiefenfazies zu freiem SiO_2 geführt." It is also well known that a hornblendic rock without quartz may, even after complete solidification, be changed to a biotite-quartz rock during metamorphism, when there is special activity of volatile constituents, including water. An example has been described by Cushing.[3]

It is not to be imagined that any of these reactions in the magma begin abruptly at any special stage in the history of the magma. For any given concentration of the molecules $KAlSi_3O_8$

[1] "Die gasförmigen Mineralizatoren im Magma," *Geologische Rundschau*, Band III (1912), 479.

[2] *U.S. Geol. Survey, 12th Ann. Rept.*, I, (1891), 657; Iddings, *Igneous Rocks*, I, 152–53.

[3] "Geology of the Thousand Islands Region," *N.Y. State Museum, Bull. 145*, p. 40.

and $NaAlSi_3O_8$, however small, there is a certain corresponding concentration of $KAlSiO_4$, $NaAlSiO_4$, and SiO_2. During the *slow* crystallization of the basaltic magma, with the continual increase in the concentration of $KAlSi_3O_8$, $NaAlSi_3O_8$, and the promoting agent water, there is a corresponding increase in the concentrations of $KAlSiO_4$, $NaAlSiO_4$, SiO_2, and others, until finally some of these exceed their saturation limit. SiO_2 then separates as quartz; $KAlSiO_4$ with $HAlSiO_4$, certain complex ferromagnesian molecules, and a limited amount of $NaAlSiO_4$ separate as a solid solution making up the mineral biotite. The molecules which separate are not necessarily the most concentrated; certain others may be much more concentrated, but correspondingly more soluble. Neither is it necessary that the molecules separate in the stoichiometric proportions represented by the reactions given. They are formed in the liquid in these proportions, but the extent of their separation from the liquid is determined by their solubility relations. The relative amounts of quartz and biotite contained in a given rock have, therefore, no necessary relation to the proportions of the various molecules indicated in the foregoing reactions. Indeed, quartz, the one product of these reactions, usually is greatly in excess of biotite, a grouping of several of the other products. This means that there is a relative storing up in the liquid of the molecules, other than SiO_2, which result from these reactions. The molecules stored up are, then, principally the aluminous orthosilicates of the alkalis, a fact of great significance in its bearing on the origin of nephelite syenites.

THE CRYSTALLIZATION OF BASALTIC MAGMA

Let us now review briefly the crystallization of a magma of basaltic composition. If the magma is cooled very rapidly, it crystallizes rapidly, giving simply a diabase. If there is rather slower cooling, there may be a limited amount of sinking of crystals of the early-formed pyroxene and more calcic plagioclase, and a consequent enrichment of the uppermost liquid in alkaline feldspar, giving a diorite as a light, upper differentiate. If the cooling is still slower, there is more prolonged opportunity for the sinking of crystals and the liquid may be sufficiently enriched in alkaline feldspar

molecules and water to give a high concentration and consequent separation of some of those molecules which are formed by the breakdown of the alkaline feldspar molecules (biotite or quartz or both). Thus may originate a quartz-diorite as a salic differentiate, or with still slower cooling a granodiorite or a granite. When cooled somewhat too quickly to allow appreciable settling of crystals, zoning of plagioclase crystals may bring about the same effect on the liquid, and the salic types mentioned above may occur as interstitial material.

From some basaltic magma the early formation of olivine may take place, with the result that the amount of possible salic differentiate is augmented. This early separation of olivine may occur during the slow cooling of a magma which with quick cooling gives no olivine but only plagioclase and pyroxene. Further, from some basaltic magma, which even with quick cooling gives considerable olivine, spinel may be formed among the products of early crystallization when the cooling is slow (see p. 30). Peridotite or spinel-bearing peridotite may, therefore, be formed by collection of the early crystals and the salic differentiate be thereby increased.

The crystallization of a normal basaltic magma always tends toward biotite granite, but the biotite granite stage is reached only with slow cooling. With somewhat quicker cooling, the process of sinking of crystals cannot keep pace with the quicker using up of the liquid, owing to more rapid crystallization. Crystallization and therefore differentiation becomes complete at the granodiorite, quartz diorite, or diorite stage according to the rate of cooling. It is clear that a dioritic border phase occurring about a granite or granodiorite mass is to be attributed to the fact that cooling at the border was maintained considerably in advance of that farther away from the border, and not to any diffusion of basic material toward the cold wall. In the cooler outer portion, crystallization was complete at the diorite stage, but in the warmer interior, differentiation by removal of crystals continued till the later stage of granodiorite or granite.[1] Daly's explanation of basic

[1] B. S. Butler has come to this conclusion in the case of a great number of stocks in different sections of Utah (personal communication from Mr. Butler shortly to be published in *Economic Geology*).

border phases is therefore substantially correct. It is not, however, necessary that the border phase is the undifferentiated original magma. It is better stated in the more general form that the border phase represents a less advanced stage in the differentiation of the magma. A satellitic stock of, say, diorite occurring in connection with a granitic batholith is to be explained similarly.

The occurrence of a basic marginal phase at the top of a batholith is a reversal of the general arrangement according to density which results from the free working of the processes of differentiation. This is not due to the action of any positive process tending to produce movement of material of early separation in the opposite direction, but to the constant tendency of cooling to put an end to the gravitative process. It is to be noted that the basic border phase is a completely frozen, solid rock during the period of continuance of differentiation in the liquid immediately below it. There is no violation of the law of gravitation. If a piece of the solid rock becomes broken off, it will, of course, sink in the liquid, but only then. It is in this manner that many basic cognate xenoliths are formed. It is clear also that the salic phase may be injected into fractures in the frozen border types formed from the same magma. The observance of this phenomenon is not to be considered as indicating any considerable difference in age.

FIELD EVIDENCES OF THE CONTROL OF CRYSTALLIZATION IN DIFFERENTIATION

The foregoing deductions relative to the course of crystallization and therefore of differentiation are fully supported by field evidence. Daly's experience has led him to the same conclusion with regard to the results, though the attainment of the results is differently explained by him. Speaking of his theory he states: "It considers the final acid pole of splitting in large post-Keewatin batholiths as granite." When differentiation is "arrested midway" "rocks of intermediate composition result."[1] He rightly considers that an intermediate rock such as diorite is the result of arrested differentiation and not an anchi-eutectic rock formed by the crystallization of a *final* eutectic liquid beyond which differ-

[1] *Igneous Rocks and Their Origin*, p. 361.

entiation cannot proceed. The use of the word "splitting" indicates that Daly considers that the medium of accomplishment of differentiation is liquid immiscibility, a process of which there is no clear evidence and of which a detailed description applied to rocks has never been offered by any investigator. The facts are, however, in perfect accord with the details of differentiation by crystallization given in the present paper.

The gravitative arrangement of the various differentiates of the Moyie sills and their individual mineralogy are absolutely that to be expected if differentiation by crystallization as outlined is the sole control.[1] They vary from normal diabase through hornblende diorite to biotite granite.

Somewhat similar sills in the Gowganda Lake District of Ontario, described by the writer, have essentially the same relations. In the original paper it was considered that the surrounding sediments played an important part in the formation of the granophyric bodies at the upper surface of the sills.[2] This opinion was arrived at principally because of the difficulty of picturing any process of pure differentiation whereby a quartzose rock could be formed from basaltic magma. With this difficulty removed the writer has no hesitation in concluding that the granophyre and the micropegmatite interstices of the diabase were formed after the manner detailed in the present paper and that interchange of material between the granophyre and adinolized sediment was a subsidiary process[3] contributing to the soda-rich nature of the border phases.

The great sheetlike mass at Sudbury with granite in the upper part and norite in the lower shows the same type of crystallization differentiation controlled by gravity.

There are also many cases of this general type with the additional feature that peridotite occurs as a basal differentiate. The Duluth laccolith, closely related in age and in origin to the Sudbury mass, is a case in point.[4]

[1] Daly, *Am. Jour. Sci.* (4), XX (1905), 187; Schofield, "The Origin of Granite (Micropegmatite) in the Purcell Sills," *Geol. Survey Canada, Museum Bull. No. 2*, pp. 1–34.

[2] N. L. Bowen, *Jour. Geol.*, XVIII (1910), 658.

[3] Cf. Collins, *Geol. Survey Canada, Mem. 33*, pp. 59 f.

[4] W. S. Bayley, *Jour. Geol.*, II (1894), 814.

It has been pointed out in the discussion of crystallization that the olivine-rich portion (peridotite) would be completely crystalline while the remaining portion was still liquid and that the same is true of the gabbro portion with respect to the siliceous differentiate. If gravitative differentiation took place under perfectly quiet conditions, the relations between the types would be transitional, but any disturbance of the mass during consolidation might cause the gabbro to have an intrusive relation to the peridotite, owing to the injection of dykes into cracks, and the granite to have a similar relation to the gabbro. In the Duluth laccolith the granitic phase has this intrusive relation.

Harker describes the intrusive igneous complex of the Cuillin and Red Hills as consisting principally of three laccoliths of peridotite, gabbro, and granite respectively.[1] In view of what has been ascertained concerning the course of crystallization in artificial melts which have a direct bearing on the question of crystallization of basaltic magma, the writer is emboldened to offer the following alternative suggestion. The igneous mass referred to may be, for the most part, one laccolith with a general eastward dip, formed as the result of a single *principal* act of intrusion of magma of the same composition as that extruded in colossal quantity during the fissure-eruptions. Rapid cooling of this magma in the flows gave basalts rather unusually *rich in olivine*. Slow cooling of the same magma in the laccolith, accompanied by sinking of crystals, would give, as we may deduce from experimentally ascertained facts, *spinel-bearing peridotite*[2] toward the western margin or base, gabbro at intermediate levels, and granite toward the eastern margin or top, the granite itself becoming more acid eastward, as Harker has observed. The intrusive relation of the gabbro against peridotite and of granite against gabbro would then be referred to movements during consolidation, of which there is abundant evidence.

[1] "The Tertiary Igneous Rocks of Skye," *Mem. Geol. Survey United Kingdom*, 1904.

[2] It is possible that even the common types of basaltic magma, not especially rich in olivine, will give spinel as an early precipitate if very slowly cooled, for in the reverse process—slow heating of the Palisade diabase of New Jersey—Merwin has witnessed the formation of a spinel-like mineral. Cf. Sosman and Merwin, "Data on the Intrusion Temperature of the Palisade Diabase," *Jour. Wash. Acad. Sci.*, III (1913), 392.

The quite exceptional degree of interaction between any two types when one is injected into the other during these movements is due to the fact, to be expected on the present supposition, that the invaded rock, though solid, was still hot, as Harker has supposed it to be. The marked amount of solution of the peridotite by the gabbro is simply the same reaction that would have happened between the liquid and the individual olivine crystals had they not collected into a body. The remarks made on p. 31 with reference to the eruptibility of peridotite to form separate intrusions of that rock would presumably apply to this natural example.

The igneous complex of the Nischne-Tagilsk region in the Urals is intrusive into Devonian sandstones, limestones, and basic lavas. As mapped by Wyssotsky[1] it appears to be, like the Iss complex, a concordant intrusion and suggests a sheetlike mass, dipping eastward, with the important platinum-bearing peridotite and pyroxenite near its base and, passing eastward (upward), successively gabbros, diorites, quartz diorites, and finally granites extending beyond the region mapped. From the present point of view the pyroxenite bodies occurring as borders about the dunites (collected olivine crystals) are *huge reaction-rims* formed in the same general manner as the borders of pyroxene seen about individual olivine crystals in a great variety of rocks (see Fig. 9).

To the writer it seems that some of these examples, especially the Skye complex, form an excellent transition type from those in which movements have been practically absent and the relation between the various rock-types is transitional, through those in which the transitional relations are destroyed and even the simple gravitative arrangement is on the verge of destruction as in the Skye example, to those in which the evidence for both of these has been obscured and for which the gravitative control of differentiation is, therefore, not demonstrable. The Cortlandt series of New York state which presents types more or less parallel to those described by Harker though richer in varieties may, perhaps, represent the last condition.[2]

[1] *Mém. Comité Géol. de la Russie*, Nouv. sér., livr. 62, 1913.

[2] G. H. Williams, *Am. Jour. Sci.* (3), XXXIII (1887), 33; G. S. Rogers, *Ann. N.Y. Acad. Sci.*, XXI (1911), 11–86.

Igneous complexes consisting of dykes, sheets, flows, and bodies without visible floor quite often show the same general association of types as that pointed out in the foregoing, and even with

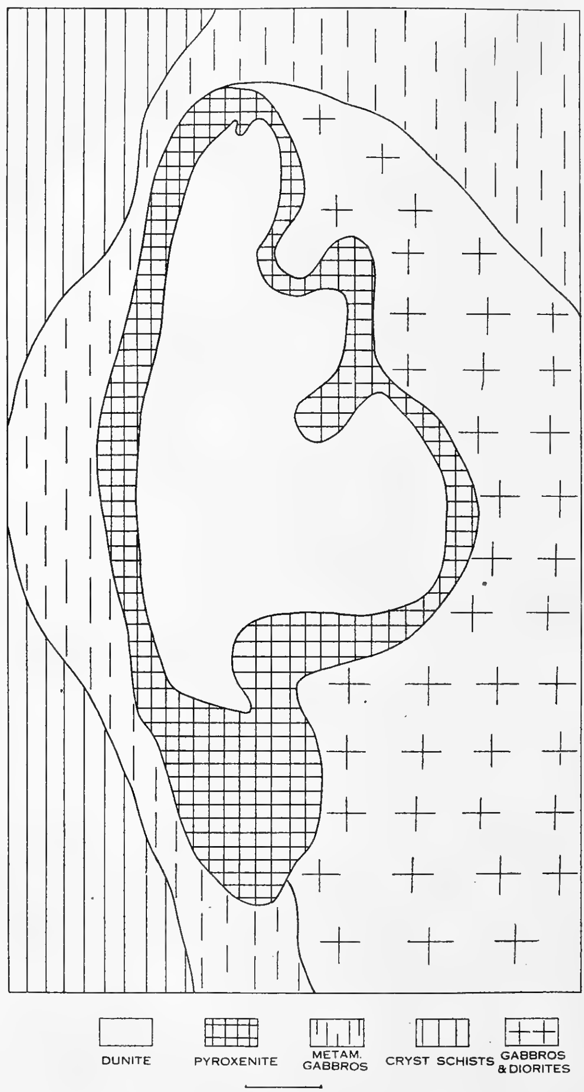

FIG. 9.—Map of the Tagilsk platinum center, Ural Mountains, after Wyssotsky, *op. cit.*, and Duparc, *Arch. Sci. Phys. et Nat., Genève*, XXXI (1911), 216.

these the control of gravity may be reasonably inferred, though the heavier differentiates may occur at unobservable depth in the case of the larger intrusions. In the following a few such bodies are described and in some cases the facts on which the foregoing inference may be based are stated.

The Hope batholith, a late portion of the Coast Range batholithic complex sectioned by the Fraser River Valley, illustrates differentiation on a huge scale, as does its southward continuation the Chilliwack batholith.[1] The types represented vary continuously from gabbro-diorite to granite, and the mineralogy of the individual types is instructive in this connection. The gabbro-diorite is a labrodorite-hypersthene rock. As the plagioclase becomes more sodic, hornblende replaces hypersthene, at first partially and finally completely. Biotite then begins to take the place of hornblende, but only in those phases which have also free quartz, and as these two appear, orthoclase also becomes important. Finally when quartz and orthoclase are fairly abundant, biotite becomes the dominant colored constituent.[2]

Precisely the same type of differentiation is shown in the plutonic complex in the vicinity of Hedley, the variation being from gabbro to biotite granodiorite. The relations at this place are such as to indicate that assimilation had no part in the formation of the salic types. The small and quickly cooled satellitic stocks and sills of gabbro and the diorite-gabbro phase occurring as a chilled border, representing restricted differentiation, about the somewhat larger quartz diorite stocks seem to show that the original magma when it arrived at its present level was still normal gabbroic magma and had not been acidified. Crystallization and the settling of crystals from this magma gave rise to the diorite in the stocks and, in a more advanced form, to the granodiorite of the batholithic mass.

The dioritic intrusives of the Prince of Wales Islands, Alaska, satellitic to the Coast Range batholith, show this same tendency

[1] R. A. Daly, "Geology of the North American Cordillera at the Forty-ninth Parallel," *Geol. Survey Canada, Mem. 38*, p. 534.

[2] N. L. Bowen, *Guide Book No. 8*, Part II, XII[me] Congrès Géologique International, 1913, p. 258, mentioned there under the field designation "hornblende-rich granitic rocks" on account of the dominance of the intermediate types.

toward a more basic nature of small bodies as compared with the more acid nature (granodiorite and granite) of the larger bodies of the main range.

The same may be said of the gabbro-quartz monzonite series of the Boulder batholith and its satellitic stocks in Montana. Again, in the Sierra Nevada, as represented in the Pyramid Peak area described by Lindgren,[1] a series of the same type is shown.

The relations of the various types of the great igneous complexes of the Coast Range and of the Sierra Nevada seem to the writer strongly to favor the opinion that all of the magma concerned in the formation of the deep-seated bodies arrived practically in the position in which we find these bodies, as *basaltic magma*. Where the magma occurred as a small body it crystallized quickly to a diabase or gabbro. This gabbroic phase may, moreover, occur as a marginal phase about a larger body whose main visible portion is diorite, the diorite being formed, *practically in place*, by the sinking out of crystals from basaltic magma. In still larger bodies a more prolonged period of sinking out of crystals has left the main visible portion a quartz diorite or a granodiorite, again formed in place, with perhaps a border phase of diorite where the process has been somewhat restricted. The magma may, of course, be injected into the surrounding rocks at any late stage and form there small dykes, etc., of the more acid types, but this fact should not be allowed to obscure the plain tendency of the acid rocks to occur principally as large masses and of the basic rocks to occur exclusively as relatively small masses.

Since the granodiorite magma comes into being only by virtue of the very slow cooling of the larger bodies, it is still liquid long after the diorite is a solid rock, and the same is true of diorite magma with respect to gabbro. Such disturbances as may occur during the consolidation of the complex may therefore cause the granodiorite to acquire an intrusive relation against the diorite and the diorite against gabbro. These relations are not to be considered as indicating the order of arrival of magmas from the depths, the order being due to a supposed primary or deep-seated differentiation. The observed sequence is to be explained rather on the basis of a

[1] *Am. Jour. Sci.* (4), III (1897), 301.

single, principal act of intrusion of basaltic magma with subsequent differentiation practically in place and controlled in a general way by the size and rate of cooling of the individual bodies formed. The sequence of intrusion observed is the result of a sequence of consolidation[1] and of movements, resulting in a certain amount of injection of one type into another, which may be wholly of a minor nature and quite subordinate as compared with the main act of intrusion of basaltic magma. In some localities more than one principal intrusion of basaltic magma took place, and therefore two or more interlocking gabbro-granite sequences may exist, with resultant irregularities in the order of succession.

The wholly random spatial arrangement of granite, granodiorite, monzonite, etc., often observed in the larger bodies plainly cannot be the result of differentiation alone, whatever kind of differentiation one may have a leaning toward. The arrangement is the result of movements following differentiation and should not, by itself, be made the basis of either affirmation or denial of gravitative differentiation. Nevertheless, the differentiation is of the type which in other cases is clearly due to a separation of liquid from crystals (fractionation),[2] either by the sinking of crystals or by the squeezing out of residual liquid, a fact which is sufficient to establish a strong presumption in favor of these processes even when they are not demonstrable.

THE ALKALINE ROCKS

NEPHELITE SYENITES AND RELATED TYPES

Biotite granite is, however, not necessarily the end or pole in the process of differentiation which has been outlined. The process may end and, very often, does end at this stage just as it sometimes ends at the granodiorite or diorite stage if the cooling is at the proper rate. Indeed, if magmas were anhydrous, we may conclude from the information furnished by the investigated systems that a granite would be the necessary end beyond which differentiation could not proceed; not a biotite granite, however,

[1] See Ussing, *Geology of the Country around Julianehaab, Greenland*, p. 312.

[2] See A. Knopf, *U.S. Geol. Survey, Bull. 527*, p. 35, on the origin of aplites associated with quartz monzonite of the Boulder batholith.

but rather one characterized by diopsidic pyroxene. But such granites are rare—many times less common than hornblende granites, and these in turn many times less common than biotite granites. The formation of the molecules which enter into the make-up of biotite results from various equilibrium reactions in which water plays an essential rôle. If any of the products of these reactions are removed from participation, through continuance of the processes of separation of liquid from crystals, further differentiation beyond the biotite granite stage is thereby permitted.

It has been shown that at the biotite granite stage, and to a lesser extent in preceding stages, reactions take place in the liquid whose principal feature is the breakdown of polysilicate molecules, probably under the influence of water, to the simpler orthosilicate molecules, among them $KAlSiO_4$ and $NaAlSiO_4$. The precipitation of $KAlSi_3O_8$, $NaAlSi_3O_8$, $KAlSiO_4$ in mica, and SiO_2 as quartz means the concentration in the liquid of all the other molecules indicated in the reactions given on pp. 44 and 45. These are principally $NaAlSiO_4$ and the volatile constituents, water, chlorine, etc., with their compounds. If the crystals of the biotite granite stage, including quartz, sink out of this liquid,[1] then the concentration of $NaAlSiO_4$ will finally reach a stage where nephelite will begin to precipitate. There may also result a concentration of CO_2, S, SO_3, Cl, etc., sufficient to cause their precipitation in compounds such as cancrinite, lazurite, hauynite, and sodalite, minerals which are peculiar to nephelite syenites and related rocks.

The field association of many nephelite syenites and their mineralogy point clearly to their crystallization at this stage of the magma. Both the mineralogy of the rocks, and the high fluidity of the magma indicated by quick changes of composition in alkaline rocks point to the high concentration of volatile constituents at this stage.

Differentiation during crystallization from these very fluid magmas will take place very freely and the formation of both

[1] Or, as seems more likely, if squeezing out of residual liquid is accomplished at the appropriate stage. The appropriate time may, perhaps, be that immediately preceding the period of marked resorption of quartz which is well attested in most rhyolites. A very important effect on the future course of the liquid would ensue, analogous to that which results if crystals and liquid are separated when olivine is about to be resorbed.

highly femic and highly alkalic types (rich in feldspathoids) may result. The more "basic" types of alkaline rock are not, however, in all cases basic differentiates from nephelite syenite magma. The reactions preliminary to the separation of quartz and biotite begin at an early stage in the crystallization of basaltic magma, and the separation of these minerals may take place at an early stage, giving rise to quartz diorites or granodiorites. The possibility of the formation of alkalic magma at a stage much earlier than the biotite granite stage is thereby introduced if conditions are favorable. Favorable conditions seem to consist in the opportunity for sinking, not only of the plagioclase crystals and femic minerals, but also of quartz crystals in sufficient amount. Thus may result relatively "basic" alkaline magmas from which such rocks as basanite might be formed, and nephelite syenite itself as a light differentiate.

In passing from the normal biotite granites direct to the nephelite syenites and related rocks, an important stage has been passed over, viz., the stage of the alkaline granites and syenites. Many occurrences, to be referred to later, show that alkaline granites and syenites are intermediate in their time relations between normal biotite granite and feldspathoid-bearing types. These facts find a likely explanation in terms of the general system here outlined. The precipitation from the granite magma of relatively large amounts of potash molecules as orthoclase and of both potash and magnesia as biotite to form normal granite results in a relative concentration of sodic and iron molecules which are precipitated as relatively large amounts of albite and as the alkaline pyroxenes and hornblendes characteristic of alkaline granites, i.e., of this state of precipitation.

In north-central Wisconsin the sequence described by Weidman—gabbro, diorite, normal granite, soda granite, quartz syenite, nephelite syenite—strictly accords with the outline of differentiation offered.[1]

The sequence in Essex County, Massachusetts, studied in great detail by Clapp, might be described substantially by repeating the type names used above in connection with the Wisconsin

[1] *Wis. Nat. Hist. Survey, Bull. 16*, pp. 349 f.

area.[1] It may be noted, moreover, that a broad study of the granites of eastern Massachusetts and Rhode Island, a region which includes the Essex County area, has led Loughlin and Heckinger to the conclusion that the alkaline granites (riebeckite- and aegirite-bearing), though later than the biotite granites, are merely the last product of the general magma represented dominantly by the normal biotite granites[2] and are not separated from them by any important period of time.

A parallel succession, viz., from basic through quartzose to feldspathoid-bearing types, appears to obtain in the post-Cretaceous minor intrusions of the Black Hills of South Dakota.[3] Continuous passage from normal granite through syenite to nephelite syenite is shown in the Bancroft area of Ontario.[4] In this case, typical alkaline granite is suppressed. According to Loewinson-Lessing, the connection of the miaskite (nephelite-syenite) of the Urals with granite has been shown by Beljankin.[5] Kerr concludes that the nephelite syenites, quartz syenites, and other alkaline syenites of the Port Coldwell area are "merely a peripheral differentiation phase of the fundamental gneiss" (granite).[6]

It should be admitted, however, that much is yet to be accomplished before one can speak with much assurance of the physical chemistry of the problems connected with the interrelations of the alkaline rocks. It will probably be a long time before important aid in attacking the questions can be expected from the experimental side, on account of the difficulty of treating systems containing volatile components. Nevertheless, it is clear that the alkaline rocks belong to a stage of great concentration of these volatile constituents and that many of the reactions in the magma depend upon this concentration. The tendency of the increased abundance of these substances, including water and the various mineral acids, is to

[1] C. H. Clapp, *The Igneous Rocks of Essex County, Massachusetts*, abstract of thesis, Massachusetts Institute of Technology, 1910.

[2] *Am. Jour. Sci.* (4), XXXVIII (1914), 55.

[3] J. P. Iddings, *Igneous Rocks*, II, 407.

[4] Adams and Barlow, *Geol. Survey Canada, Mem. 6*, 1910, p. 260.

[5] F. Loewinson-Lessing, "Origin of the Igneous Rocks," *Geol. Mag.*, N.S., Dec. V, Vol. VIII, p. 255.

[6] *Ontario Bur. Mines, Ann. Rept.*, 1910, p. 230.

displace silica from its combinations. Thus is brought about great concentration of the molecules which go to form biotite at the stage of the normal granite, and biotite, together with quartz, is precipitated. From similar causes results the precipitation of the olivine fayalite in the alkaline granites, the felspathoids, and even free alumina in foyaitic rocks. If the question is attacked from this point of view, that is, if the various minerals precipitated at any stage are recognized as those least soluble (in proportion to their abundance) at that stage, if the various equilibrium reactions indicated by these minerals are sought out, and if the effect on these reactions of the removal of minerals is considered, it seems likely that a good understanding of the alkaline rocks might be arrived at from field studies with a certain amount of guidance afforded by experimental work.

Discussion of the question whether alkalic and sub-alkalic rocks may be derived from the same magma has not been included in the foregoing because it is believed that field facts leave no possible doubt in the matter.[1] Crystallization proceeds, apparently, with the precipitation of the constituents of normal granite, then of those of the alkaline granites, through those of the alkaline syenites, to those of the nephelite syenites, with no sharp lines of demarkation anywhere in the series. It is to be remembered that the series will be maintained only when the continual removal of crystals from that part of the liquid in which they formed is likewise maintained.

The genetic relation and continual passage from sub-alkaline to alkaline types is now recognized as true by most petrologists and is, in fact, well illustrated by the difficulty encountered by every petrologist in setting up a boundary between them. There appears to be, however, a peculiar subconscious persistence of the older idea of a genetic separation even in the minds of many who admit a genetic connection. This tendency is apparent especially in nomenclature. A perfectly normal diabasic or gabbroic rock, for example, is likely to be termed an essexite if it occurs in a region of dominantly alkaline rocks and especially if associated with true

[1] See H. S. Washington, "The Volcanic Cycles in Sardinia," *Compte-rendu,* XII^{me} session, Congrès Géologique International, p. 235.

essexite. Such a tendency in naming rocks is to be guarded against, for it covers up facts—in this case the fact of frequent association of normal gabbroic rocks with alkaline types.

LEUCITE-BEARING ROCKS

Rocks bearing leucite are of rare occurrence and their relations are, perhaps, not well enough understood to make clear the manner of their origin. Leucite is sometimes one of the minerals precipitated from a nephelite syenite magma derived, presumably, in the manner already outlined. The relations of some leucite-bearing rocks suggest, however, that differentiation in the magma from which they were formed proceeded along the lines outlined in the discussion of the formation of biotite and that the magma was then intruded to a high level where water could escape more or less freely. The lowering of the pressure of water-vapor lessens the possibility of the formation of those hydrous molecules which go to form mica, and a part of the potash must then be precipitated in some form other than mica, viz., as leucite.[1] Highwood Peak stock, for example, sends out mica-rich dykes, and these, where they cut loose breccias which would facilitate the escape of water, often pass in a single dyke into a leucite-bearing rock, so-called leucite basalt.[2] Iddings has called attention to the fact that some minettes and some leucite basalts have nearly identical composition. Shonkinite, a type comparatively rich in biotite, has no mineralogical equivalent among the effusive rocks, but is represented by its chemical equivalent, which is a leucite rock, owing presumably to the loss of water on extrusion. It seems possible that the gradual escape of water from the outer border of the Shonkin Sag laccolith has given rise to the 5–15-foot border of leucite-basalt which is nearly identical in composition with the shonkinite immediately adjacent to it.[3] If the partial escape of water takes place from a moderately deep-seated body, even though not especially rich in the molecules which go to form mica, it seems possible that further

[1] Cf. H. S. Washington, "The Formation of Leucite in Igneous Rocks," *Jour. Geol.*, XV (1907), 377–79.

[2] L. V. Pirsson, *U.S. Geol. Survey, Bull. 237*, 1905, p. 22.

[3] Pirrson, *Op. cit.*, p. 47.

differentiation, with leucite as one of the crystalline units in that process might give rise to rocks comparatively rich in leucite. The relative abundance of leucite-bearing types as effusive and hypabyssal rocks as compared with their rarity as truly abyssal rocks lends support to the idea that near-surface conditions favor their formation. Such considerations raise the whole question of the extent to which evaporation, on the one hand, and cooling on the other, have controlled the crystallization of a given rock mass.

Under the foregoing conception of the leucite rocks they should presumably belong at a late stage of the genetic sequence, a stage appropriate to the preliminary formation of biotite. This appears to be true of the lavas of the Aeolian Isles as determined by Bergeat, who refers to the leucite types as an old-age manifestation following after a sequence exhibiting increasing acidity.[1]

ESCAPE OF THERMAL WATERS

The crystallization of the constituents of the nephelite syenites and related rocks is not the end of the process of crystallization from the magma. The precipitation of analcite and other zeolites, apophylite, and thomsonite follows thereafter and the process passes on into the stage considered to belong rather to that of thermal waters than to magmas. In these waters there is an increasing concentration of the very soluble salts of the alkalis such as Na_2S, Na_2CO_3, NaCl, etc. (see equilibrium reactions, pp. 44 and 45), and they probably constitute the alkaline ascending waters of primary ore deposits. These waters may, of course, be expelled from the magma at a much earlier stage, say the granitic or granodioritic, if the cooling is at such a rate that differentiation is completed at that stage.

OTHER CONCEPTIONS OF THE ALKALINE ROCKS

Daly has offered a theory of the alkaline rocks in general, which supposes that they are differentiates from subalkaline magmas that have absorbed limestone. The formation of feldspathoids is considered due to a desilication of the magma consequent upon the

[1] A. Bergeat, *Compte-rendu*, XII^{me} session, Congrès Géologique International, Canada, p. 250; "Die äolischen Inseln," *Abhand. der k. bayer. Akademie der Wiss.*, II, Kl., XX, (Munich, 1899), p. 270.

introduction of lime and its subsequent removal as lime silicates during the course of differentiation. The occurrence of cancrinite and primary calcite is considered evidence of the introduction of CO_2.

The fact that limestone may, as a rule, be found in the same general region as alkaline rocks cannot, in itself, be regarded as having any significance, because limestone is of such extremely widespread occurrence. Even in those cases which seem best to indicate some connection between the limestone and the alkaline rock, the application of the theory presents grave difficulties. In the Haliburton-Bancroft area of Ontario, batholiths, dominantly granitic, show nephelite syenites against bordering limestones. The relation is striking, but since the theory demands the absorption of limestone by the magma, and the subsequent formation of the alkaline rock by differentiation from the homogeneous magma, the manner of differentiation should determine the location of the alkaline rock, and the fact that this differentiate borders against the limestone offers no real support to the theory. It is, moreover, clear that the magma has not been desilicated for huge volumes of it have crystallized to granite, unless the desilication was entirely local, a possibility which does not seem to accord with the idea that the limestone was absorbed as blocks which had sunk in the magma to great depths.

The writer has considered an alternative hypothesis which still maintains some connection between the limestone and the alkaline rock. The suggestion is that silica was subtracted from the magma locally for the formation of amphibolites from the surrounding limestone, a local desilication without important additions of lime from the limestone.[1] The occurrence of cancrinite could then be considered as due to introduction of CO^2 from the limestone, but this supposition, like Daly's, does not offer an explanation for the special abundance of sodalite in these nephelite syenites. There is no reason why an abundance of chlorine should be produced by interaction with limestone.

On the other hand, if it is supposed that the nephelite syenite belongs to the stage of great concentration of the volatile con-

[1] The quite similar suggestion recently offered by Foye appeared too late to receive adequate recognition in this paper, *Am. Jour. Sci.* (4), XL (1915), 430.

stituents of the magma, it is to be expected that the minerals precipitated would give evidence of richness in those volatile constituents which are known to be constituents of magmas. The minerals are, in the present case, sodalite and cancrinite, in other cases noselite and hauynite as well. Moreover, under this theory, this light residual magma tends to accumulate, during the course of differentiation by crystallization, in the upper and outer parts of the batholithic chamber, and in a limestone terrane it is especially likely to border on the limestone. This theory, then, assigns to the limestone no essential action in the formation of the alkaline types as is obviously true in other localities where limestone is absent in the country rock or present only in relatively small quantity. It is, for example, difficult to imagine how an occasional bed of limestone in a terrane consisting principally of siliceous gneisses could be considered a possible agent of desilication, for its absorption would entail simultaneous absorption of a much greater quantity of relatively siliceous material.

To return to the idea of the desilication of the magma, either localized or otherwise, through interaction with limestone in the Bancroft area, it may be pointed out that there are in this area many bodies of diorite of no inconsiderable size which one would expect to have suffered similarly, for they also have caused amphibolitization, etc., of the limestone. Indeed, the task set the limestone would seem to be much less arduous in this case than in the case of granitic magma. The reduction to nephelite of the polysilicate molecules represented in the alkaline feldspars would need to be preceded in the granitic magma by the binding of the quartz, whereas this would be unnecessary in the diorites. Nevertheless the diorite bodies apparently fail to show borders of nephelite-bearing rocks. They belong, not with the diorites, but with the quartz-rich rocks, the granites, as they should under the present conception of their origin.

Neither is the Bancroft area the only one that exhibits this intimate connection of the nephelite-bearing rocks with granites. Several cases have already been referred to in discussing the sequence of types and several more could be added. At this point only three others will be mentioned. The nephelite syenites

(canadites) and umptekites of Almunge, Sweden, are transitional both into alkaline granites and nordmarkites and also into normal sub-alkaline granites.[1] In the Bushveldt laccolithic complex, the succession, ultra-basic types, norites, granites, nephelite syenites is developed and Brouwer emphasizes a significant feature in this connection when he refers to the formation of "eenerzijds zeer kwartsrijke, anderzijds veldspatoïdvoerende gesteenten" by "dir-recte differentiatie." As a third example, it may be pointed out that Daly has himself detailed the evidences of a "genetic bond" between the Kruger nephelite syenites, malignites, etc., and the Similkameen granite.[2]

Normally, the nephelite syenites and their relatives appear, then, to fall naturally into a differentiation sequence, being intimately related to quartzose types, as the field evidence indicates and as chemistry would lead one to expect. It may, nevertheless, be reasonably considered probable that some nephelite rocks, especially melilite-nephelite types, may be formed by the method suggested by Daly.

Harker[3] has advanced the idea that the alkaline rocks owe their origin to a certain type of earth movement distinct from that brought into play during the formation of sub-alkaline types. Harker admits that the connection is not understood, but seeks to show that the distribution of the two branches is related to the great tectonic features of the earth. So many exceptions have been pointed out by Cross, Daly, and others to his general subdivision, with Becke, into Atlantic and Pacific provinces, that its validity will not be discussed further here. There remain, however, certain tendencies in that direction, among them the striking fact that the Rocky Mountain front in the United States lies approximately along a boundary between regions of dominantly alkaline and dominantly sub-alkaline rocks.[4] Harker has discussed this feature in some detail for the state of Montana. The general con-

[1] P. D. Quensel, "The Alkaline Rocks of Almunge," *Bull. Geol. Inst. Upsala*, XII, 157.

[2] *Geol. Survey Canada, Mem. 38*, Part I (1912), p. 459.

[3] *The Natural History of Igneous Rocks*, p. 93.

[4] Daly, however, describes nephelite syenite, etc., from the heart of the Cordillera (*Geol. Survey Canada, Mem. 38*, p. 488).

ception of the alkaline rocks advanced in the present paper appears to afford a rational explanation for this fact. It is considered that the alkaline rocks are derivatives of the same magma as sub-alkaline rocks, and belong to a late stage of concentration of the volatile constituents of the magma. They are light differentiates and therefore have a distinct *tendency* to occur as marginal and satellitic bodies *at a higher level* than their sub-alkaline co-derivatives. In those discrete mountain groups which lie east of the Rocky Mountain front in Montana, erosion has, as a rule, penetrated only to a level where the igneous intrusives are dominantly alkaline. Farther to the west, uplift and consequently depth of erosion are relatively greater; the present surface has penetrated, for the most part, below the level of the alkaline intrusives, and the principal exposures are of sub-alkaline types. Farther to the north, in the Rocky Mountain system in British Columbia, erosion has as yet penetrated only one considerable igneous body, the Ice River laccolith, and that, in conformity with the expectation of the general tendency postulated, is an alkaline body.[1]

If the suggestion is correct, then, this striking division into provinces is only in a remote manner connected with the tectonic feature, the Rocky Mountain front. Relative depth of erosion is accountable for the differences observed at the present surface.

It appears, moreover, that the general tendency toward a greater abundance of alkaline rocks in recent terranes as compared with ancient terranes is assignable to an accompanying tendency toward a greater depth of erosion in the ancient terranes.

Smythe, attacking the subject from a somewhat different viewpoint, has arrived at substantially the same conclusion as the writer with regard to the importance of mineralizers in the formation of the alkaline rocks.[2] Daly, too, appears to consider it possible that the CO_2 added when limestone is absorbed may be an important factor and this may perhaps be true in some cases. The writer would agree with Smythe in considering the production of

[1] J. A. Allan, *Geology of the Ice River District, British Columbia*, abstract of thesis, Massachusetts Institute of Technology, 1912, and *Geol. Survey Canada, Mem. 55*, 1914, p. 209.

[2] C. H. Smythe, Jr., *Am. Jour. Sci.* (4), XXXVI (1913), 46.

alkaline types entirely independent of absorption of foreign material in the great majority of cases and would assign the significant rôle to juvenile gaseous substances, principally water, concentrated in residual liquors in a natural way.

IS BASALTIC MAGMA THE PARENTAL MAGMA OF ALL IGNEOUS ROCKS?

GENERAL CONSIDERATIONS

In the foregoing pages an attempt has been made to give a rather full discussion of differentiation in igneous rocks and of the principles underlying it. A beginning was made with the discussion of the common association, in sill-like bodies where both top and bottom are exposed and all the relations clear,[1] of diabase with biotite granite and often several intermediate types. A specially notable case is that of the Purcell sills described by Daly and later by Schofield in greater detail. It was shown that the relations found are strictly those to be expected if differentiation takes place through crystallization under the influence of gravity. The relation of the biotite granites and related biotite-quartz types to the alkaline rocks was then discussed and it was shown that the same principles of derivation are applicable.

While no implication is intended in the foregoing that basaltic magma is always the parent magma of the igneous series given, and while the outline may equally well be regarded as a contribution to the study of the differentiation of dioritic magma or of granitic or other magmas, nevertheless the discussion does emphasize the fact that a great variety of igneous rocks, possibly all important types, could be derived from basaltic magma by differentiation under the appropriate conditions. Since the idea that basaltic magma is the parent magma of all igneous rocks has already found favor with some petrologists, it becomes advisable to discuss the matter in the present connection.

Of the visible igneous rocks of the globe the volume of salic types is many times that of femic types.[2] In view of this fact it may seem a bold step to express the opinion that basaltic magma

[1] A method of attack the importance of which Daly has long emphasized.

[2] See Daly, *Igneous Rocks and Their Origin*, p. 52.

is the primary igneous material from which all known igneous rocks are derived, yet it is a logical step. Any large salic body, say granitic, whose every exposed portion is simply granite tells us very little about itself. Such are the bodies which give salic rocks their great preponderance. It appears to be the prevailing opinion, seldom explicitly stated, perhaps, but certainly implied in description and discussion, that such bodies of granite were formed by the crystallization of granitic magma which always existed *as such.* The opinion is not based on any definite evidence to that effect but merely on a lack of direct evidence to the contrary. Such lack of evidence stands over against a mass of convincing and positive evidence of the origin of granites (and other salic types) which occur in bodies of such size (or shape) that their relations are clearly shown.

A broad survey of the Keweenawan eruptives of the Temiskaming–Lake Superior region lays bare the essential features of the differentiation of basaltic magma as it is controlled by the size of the body. The Keweenawan flows are overwhelmingly normal basalts. The quickly cooled smaller dykes and sills are principally normal diabase, sometimes olivine diabase. The sills of moderate thickness have diabase with micropegmatite interstices as a result of somewhat slower cooling. The sills of considerable thickness commonly have diabase with micropegmatite interstices and a small salic differentiate (with intermediate types) at their upper borders. The writer has seen several examples both in the sills of Gowganda Lake and in the Logan sills of the north shore of Lake Superior.[1] In the case of the huge sheetlike mass at Sudbury the salic differentiate at the top constitutes a considerable area of granite. It is fortunate for the geologic investigator that the Sudbury intrusive sheet lies in a basin-shaped structural depression. If it had assumed the form of an inverted basin, erosion might have exposed the granitic phase only and the derivation of the granite by gravitative differentiation would not have been revealed. It would then have passed muster as an ordinary granite batholith, and, no doubt, many batholiths (bodies without visible floors) are similar to the

[1] N. L. Bowen, Gowganda Sills, *Jour. Geol.*, XXVIII (1910), 658; Logan Sills, *Ontario Bur. Mines, Ann. Rept.*, 1911, p. 127.

Sudbury sheet, in method of intrusion, great horizontal extension, and stratiform arrangement of their differentiates, of which we see only the more salic. If in any case, basaltic magma collected in a chamber comparable in size with that occupied by the Sudbury sheet, but of considerable vertical instead of horizontal extension, it is unlikely that any amount of later disturbance or erosion to be reasonably postulated would bring to light anything but the light salic differentiate. How much more true is this of bodies of magma of very great dimensions! While, therefore, nothing in the foregoing forces the conclusion that great granite bodies are merely the light differentiate of a basaltic mass, nevertheless the conclusion is entirely consistent with the observed tendency which accompanies increasing size.

The features pointed out for the North American province described are not local. They are world-wide in their occurrence. Certain observations by Sederholm are pertinent in this connection. Speaking of the fundamental rocks of Finland, he says that abyssal rocks corresponding to almost every formation of effusive rocks of the granite family are easily found in the deeply eroded complex, but no *abyssal* rocks are found belonging to the diabase family, though rocks of this latter family have been erupted during every time of quiet sedimentation in Fenno-Scandia, either forming effusive beds or being intercalated between the strata of the sediments. He considers that the erupted diabases represent basic marginal phases of the common magma from which all the rocks were derived and that they were drained off before the acid portions because of their greater fluidity.[1] It is necessary under this assumption to imagine a marked differentiation of the magma while still liquid, and an arrangement of the differentiates in opposition to gravity. It is also necessary to imagine that in all cases the basaltic marginal phase happened to be drained off. The difficulties involved are entirely avoided if it is supposed that the original magma was basaltic (diabasic) and that where intruded in small bodies it formed, for the most part, sills and flows of undifferentiated diabase or basalt. Where it collected in large abyssal masses crystallization took place slowly. Crystals sank

[1] *Bull. Comm. Géol. de Finlande*, XXIII (1907), 108.

gradually as they grew in the liquid. The liquid passed through all intermediate compositions toward the granitic and fed at various stages the dykes and flows of the different types of porphyry, naturally later than the diabasic types. But in the *visible upper portion* of the very large bodies of the magma, only those products which were precipitated at the latest (and lightest), stage of the magma's history are represented, i.e., the constituents of the rocks of the granite family.

Moreover, it seems quite reasonable and consistent to believe that the descent of igneous rocks has not been different from that outlined above, even in the earliest stages of the earth's history of which we have any knowledge from actual rock outcrops. The rocks described by Sederholm belong among the oldest. In the Canadian shield also the colossal batholiths of the Laurentian were preceded by the basaltic outpourings of the Keewatin, and the same interpretation of their relation applies as that just given for the Scandinavian shield. In the basement complex of Rhodesia the contrasted volume of exposed acid and basic types[1] is attributable to the action of similar processes. It seems possible, then, to consider that basaltic magma is the original material of all igneous action of which we have any direct knowledge, and that all igneous rocks, even the huge areas of pre-Cambrian granitic rocks are differentiates of basaltic magma.

The objection might be raised that, in the more deeply eroded terranes such as some pre-Cambrian areas must be, erosion should have cut through the salic types and exposed their complementary rocks if the salic batholiths are really only the upper portions of bodies formed by differentiation of basaltic magma. In other words, one would expect to find areas of basic rocks of batholithic dimensions. The objection seems, however, to lose much of its force when carefully considered. The granitic differentiate of an originally basaltic magma may be as much as 10 or 15 per cent of the total differentiated mass.[2] Erosion would, therefore, have to penetrate 10 or 15 per cent of the depth of the igneous mass to

[1] F. P. Mennel, *A Manual of Petrology* (London, 1913), p. 105.

[2] Collins has, for example, estimated the aplitic portions of the coarser Nipissing intrusives as 10 per cent or more of the total (*Geol. Survey Canada, Mem. 33*, 1913, p. 65).

pass the granite. Presumably this has been accomplished in many cases and somewhat more basic rocks, granodiorite or diorite, have been encountered. There is an abundance of diorite in the "granitic" areas of the pre-Cambrian and possibly careful structural and petrographic work would reveal that this diorite underlay granite. This has already been shown to be true in the case of the Palaeozoic Saugus batholith of Massachusetts.[1] Seldom has sufficiently detailed work been accomplished to reveal such facts in those areas of the pre-Cambrian where igneous rocks dominate. In the case of very large bodies erosion might penetrate two or more miles and then reach only types not very different from granite, say granodiorite or quartz diorite, which are not distinguished from granite in a good deal of areal pre-Cambrian work. Probably only profound erosion would reach the more basic types and it may reasonably be doubted whether erosion is ever sufficient unless the igneous mass is favorably disposed, as it was in the case of the Duluth laccolith. It is true that great depth of erosion is sometimes believed to obtain in pre-Cambrian areas, but this appears to be often merely the attributing of great power to geologic forces in the distant past. On the contrary, when the structural evidence is examined there has often been found but moderate erosion. Thus, in the Thousand Islands region Cushing has shown that some batholiths have been barely de-roofed,[2] and in the Bancroft area of Ontario there are still remnants of effusive rocks, comagmatic with the batholiths, which were formed on the surface about the time of the intrusion of the batholiths. Clearly the surface at that time could not have stood at a level greatly different from that of the present surface. These are isolated cases, of course, but infolded patches of surface volcanic rocks in the batholithic pre-Cambrian areas are common enough to indicate that possibly these terranes have not suffered excessive erosion. Apparently, then, one would not expect erosion to have exposed areas of basic rocks of batholithic dimensions even if the granitic batholiths were formed by differentiation of basaltic magma. It is tentatively suggested, however, that it may be the greater erosion of pre-Cambrian ter-

[1] C. H. Clapp, *op. cit.*, p. 6.

[2] "Geology of the Thousand Islands Region," *N.Y. State Museum, Bull. 145*, p. 43.

ranes which has led to the exposure in those areas, and almost solely there, of large bodies of anorthosite which should under the present conception of its origin be a moderately deep-seated differentiate.

Another fact which has been advanced in favor of the idea that basaltic magma is the parent of all igneous rocks is its continual recurrence in terranes of all ages in the form of fissure eruptions.[1] It occurs likewise as dykes filling fissures related to the primary structures of all terranes, which dykes are indeed often the feeders of fissure eruptions. These occurrences indicate its sudden arrival from great depth without opportunity for differentiation en route. All other types of magma when occurring in dyke form tend rather to center about areas of batholithic intrusion, the seats of differentiation.

It may, perhaps, be considered that it is the great fluidity of basaltic magma which permits it to pour out freely in the manner referred to. If this were true it would seem that basaltic magma would be the commonest material of fissure eruptions, andesitic magma somewhat less common, dacitic magma less common still, and so forth. Apparently, however, the material of fissure eruptions is too commonly basaltic to agree with this supposition. It is this fact which suggests that it is the only magma which arrives at the surface undifferentiated and that other magmas are formed from it by differentiation when it collects in batholithic masses.

Some further support for the conception that basaltic magma is the parent magma of all igneous rocks is to be found in the mineral constitution of the rocks themselves. Granitic or any closely related magma could not give a differentiate of basaltic composition by the collection of its heavier minerals. Only dioritic magma might reasonably be expected to do so at any time, but it would be remarkable if it gave at nearly all times basaltic magma of approximately uniform composition. One would expect rather a marked variation, according to the opportunity for the sorting of crystals of different densities, in all rocks more "basic" than the parental magma. This is true only of all rocks more basic than basalt. Pyroxenites, peridotites, and here we should include anorthosites also, are notably variable in the proportions of their

[1] Daly, *Igneous Rocks and Their Origin*, p. 458.

characteristic minerals. There is no evidence of that tendency toward fairly uniform relative proportions of mineral constituents such as is exhibited in basalt, diorite, and so forth.

THE AVERAGE IGNEOUS ROCK COMPARED WITH THE PARENTAL MAGMA

The fact that the average composition of igneous rocks is not basaltic might, perhaps, be considered to preclude the possibility that basalt is the parental magma of all rocks.

Several calculations have been made of the average igneous rock. The calculation is essentially the averaging of existing rock analyses and gives only the average analyzed rock. Even if weighted according to the relative exposed abundance of types, it is to be noted that the figures obtained would represent only the average exposed rock. It should not be assumed that this average would represent the composition of the general magma from which all rocks were derived. Such an assumption ignores the evidence of such bodies as the Sudbury sheet, the Duluth laccolith, and many others which tell us that the upper part of an igneous body is much more salic than the magma from which it is formed, *especially if it is a large mass*. Consideration of this fact with reference to large salic bodies of which only the upper portion is visible makes it necessary to believe that the salic rock is only a light differentiate of the magma from which it formed. This nearly constant tendency of the exposed rock to differ from the magma from which it formed in a special direction makes the average of the exposed rocks differ from the general magma in the same direction. There is therefore nothing contradictory to known evidence in the statement that, though the average of analyzed rocks is tonalitic-dioritic, and a weighted average would be even more acidic, yet the general magma is more basic, viz., basaltic.

The same considerations apply to the average rock of any particular area. If this average is considered to represent the original magma, and if the composition of any large body of which only the upper part is visible enters into the average, the assumption is implicit in the result of the calculation that the whole of this body is identical with its upper portion. This assumption is unsafe. The earliest of the effusive rocks

which initiate the volcanic phase of the igneous activity of the region in question is more likely to represent the original magma. In the Christiania region of Norway, for example, the average composition of the rocks is somewhere near nordmarkite. If it is considered that all the rocks were derived from nordmarkitic magma, it is necessary, in order to explain the sequence of types, to imagine a differentiation while the magma was still liquid, as Brögger does, and a collection of the heavier basic differentiates toward the top. This conception Harker has rightly described, in a different connection, as "wholly chimerical,"[1] but one might refer in a similar manner to the deep-seated magma basin, stratified according to density, as it might be, but from which, *in the usual case,* "the *earlier* intruded magmas are drawn from the *lower* levels."[2] On the other hand, if it is assumed that the original magma was the gabbro which forms the earliest extrusives, the rocks of the Christiania region seem explicable as the result of gravitative differentiation increasing in importance with the size of the bodies and emphasizing the alkaline and salic types in the exposed portions of the larger bodies. In the very closely related Julianehaab region Ussing has called special attention to "the contrast between the mode of occurrence of the dense, gabbroid rocks and the specifically lighter, more acid rocks," the former as dykes of small volume and the latter as batholiths, or rather the upper part of batholiths.[3]

THE EARLIER STAGES OF IGNEOUS ROCK EVOLUTION NOT REVEALED

The facts which have been pointed out in favor of considering basaltic magma the parent of all igneous rocks do not, of course, definitely prove the case, yet they appear to establish a strong presumption in its favor. The geologic record makes it necessary to imagine a source of basaltic magma beneath all parts of the surface of the earth throughout recorded time, and a great many facts indicate the derivation of all other magmas from the basaltic. What the antecedent stages which permitted the formation of this source of basaltic magma may have been is not revealed in the record. Consideration of these early stages would lead one into

[1] *Op. cit.*, p. 317. [2] *Ibid.*, p. 330 (italics are mine). [3] *Op. cit.*, p. 308.

cosmogony, which subject it is desired to avoid in this paper. It may be noted, however, that the existence of a source of basaltic magma does not necessarily imply a liquid layer and possibly not even a basaltic layer, since it is conceivable that disseminated basaltic liquid might have been generated as occasion demanded by some sort of selective fusion of heterogeneous non-basaltic material. This suggestion is not offered as an expression of opinion, but merely as an indication that belief in the continuous existence of a source of basaltic magma does not necessarily force one to prefer either the view that the earth was once molten or the view that it has always been a relatively cold body. Until the petrologist finds in his own field some reason for advocating the one or the other view, there is little need of his expressing a preference. Neither is it necessary for the petrologist to express an opinion as to the physical condition of the earth's interior, whether magmas come from great depth or moderate depth, whether they are generated as a result of some mechanical production of heat or have persisted from a once-molten earth. He may reasonably start with magma, however produced, and with such a magma or magmas as the geologic evidence appears to dictate.

THE NORMAL LINE OF DESCENT

The facts indicating the parental nature of basaltic magma have been detailed in the foregoing. As already stated, they are not considered as proving that basaltic magma plays this rôle, but rather as constituting a strong plea for the inclusion of this idea among the petrologist's working hypotheses. Especially is this desirable when a conception is gained of a process whereby igneous-rock series can be derived from basaltic magma. This process of evolution, through the action of crystallization differentiation, has already been stated in detail, with emphasis, for the most part, on its physico-chemical aspects. Consideration will now be given to some of its geological aspects.

In the following an outline of the conception of the igneous rocks here advanced is presented. It may serve as a summary of the fuller discussion presented in the preceding pages. Being in schematic form, it is necessarily somewhat rigid and is not to be interpreted without due reference to the discussion in the text.

I

COURSE FOLLOWED BY LIQUID AS A RESULT OF REMOVAL OF CRYSTALS

basaltic
↓
dioritic
↓
granodioritic
↓ ↓
granitic (basanitic)
↓ ↓
foyaitic (foyaitic)

(The career of the liquid may end at any of these stages on account of its complete freezing and consequent cessation of removal of crystals.)

II

MATERIALS FORMED BY ACCUMULATION OF CRYSTALS (ROCKS)

Extreme sorting.—Peridotites, pyroxenites, anorthosites, and other "anchimonomineralic" rocks.

Little sorting.—Any type formed by the freezing of the liquid at any of the stages in I, in a position where there has been accumulation, great or small, of one or more of the kinds of crystals which have recently separated from the liquid.

Principal Products of the Slow Crystallization of Basaltic Magma

Increasing concentration of volatile constituents and falling temperature ↓

STAGE		
1.	(spinellids) magnesia-iron olivine	
2.	magnesia-iron pyroxene	calcic plagioclase
3.	lime-magnesia-iron pyroxene	calcic plagioclase
4.	hornblende	medium plagioclase
5.	biotite, quartz, orthoclase	medium plagioclase
6.	biotite, muscovite, quartz, orthoclase	sodic plagioclase
7.	alkalic { hornblende, quartz, or pyroxene	orthoclase, sodic plagioclase, or microcline
8.	alkalic { hornblende, iron olivine, or pyroxene	orthoclase, sodic plagioclase, or microcline
9	alkalic { hornblende, nephelite, or pyroxene	orthoclase, sodic plagioclase, or microcline
10.	pyroxenes, corundum, nephelite, sodalite	orthoclase, sodic plagioclase, or microcline
11.	pyroxenes or olivine, analcite	orthoclase, sodic plagioclase, or microcline

The preceding outline is intended to represent the usual case. The period of crystallization of any stage is somewhat later than that of the stage above it. The crystallization of a mineral placed under any stage is essentially simultaneous with that of the other minerals grouped with it. There is, of course, considerable overlapping from stage to stage; indeed, the division into stages is purely arbitrary, for the process itself is perfectly continuous.

The writer has some hesitation in presenting this very definite statement concerning crystallization, for he knows full well that it cannot be absolutely correct. Nevertheless, it is believed that if our knowledge of the subject were complete, the final correction of the statement would be principally an elaboration rather than an alteration. In presenting the foregoing scheme, the purpose is chiefly to emphasize the point of view from which the crystallization of a magma is to be attacked. The great difference should be realized between the simple group of minerals which results when a magma is crystallized quickly and the long series of products which may result from slow crystallization. The fundamental controlling influence of the plagioclase mix-crystal series must be recognized; also the fact that the minerals precipitated give clues to certain equilibrium reactions in the liquid. Ferromagnesian material in one form or another must appear at all stages of precipitation. If a chemist has a mixture of, say, potassium and sodium salts, he has yet to discover a method of cooling or evaporation whereby *all* the sodium salts can be removed from the solution so long as any solution is left.

The great lowering of the temperatures of crystallization of granitic magma as the result of the concentration of volatile constituents has often been emphasized, but the converse is worthy of equal emphasis. The relatively great concentration of volatile substances in granitic magma is the result of its low temperature of crystallization and follows as a necessary consequence of the long-preceding period of crystallization in virtue of which alone granitic magma comes into being.

VARIATIONS FROM THE STATED SCHEME OF CRYSTALLIZATION

In presenting the schematic statement of the progress of crystallization the writer does not intend to imply that this sequence is

rigidly adhered to in every case of slow crystallization of basaltic magma. The plan of presenting the subject—that of offering experimental evidence bearing on the questions and then deducing therefrom the crystallization of a magma—has led naturally to the statement of crystallization given above. This may be regarded as the normal sequence, for, considering the volume of rock types produced, it is probably the most important. Any other succession of types may be regarded as a variant from the above. In this connection it is intended to discuss what appear to be two principal variants and to suggest that the effects of pressure probably adequately explain these variants. The effects are not those of pressure *per se*, for these are probably altogether insignificant, but rather of pressure in so far as it affects the retention of volatile constituents, especially water.

The sequence gabbro-diorite-quartz diorite-granodiorite-granite-alkaline types expressed in the tabular statement given is that normal to the Mesozoic-Tertiary intrusives of the Cordilleran region of North America. The types characteristic of the pre-Cambrian intrusives of the Adirondacks and adjacent regions of Canada are rather gabbro-diorite-syenite-granite-alkaline types. If we regard the nature of the plagioclase as a measure of the stage of crystallization or the degree of concentration of the liquid residue relative to the original volume of magma, it is clear that the types of the eastern area exhibit a delaying of those reactions, characteristic of the presence of water, which result in the separation of significant amounts of biotite and quartz. It appears, therefore, that at any stage, representing a certain degree of concentration of the residual liquid with respect to the original magma, the active mass of water was less during the development of the eastern types than during that of the Cordilleran types. In any individual case it would be impossible to refer such a difference in the activity of water definitely to a difference of external pressure, but that this is the kind of effect which would result and no doubt has resulted in many cases is hardly to be questioned.

The production of quartz and biotite in the presence of water results in large part from the breakdown of $KAlSi_3O_8$ molecules. A low value of the active mass of water, sufficient to reduce this

reaction to a minimum, should considerably advance the period of precipitation of $KAlSi_3O_8$ as orthoclase.[1] This may be the principal factor controlling the essexite line of descent consisting of gabbro, essexite, augite syenite, and the alkaline types common to all lines of descent. Thus the precipitation of considerable orthoclase takes place at a stage when the plagioclase is still strongly calcic in the case of essexite, and the delaying of those reactions characteristic of great "activity" of water is still well marked at the augite syenite stage, when the feldspar is already entirely alkalic whereas the dominant ferromagnesian mineral is still a typically anhydrous pryoxene. Even in this series, however, the water finally makes its presence felt as its concentration in the residual liquid increases, with the resultant production of feldspathoid-bearing and of quartzose types.

An apparent association of rocks of the essexite line of descent with specially intensive tensional faulting (e.g., Christiania and Julianehaab) lends support to the foregoing conclusion that this line of descent is controlled by low pressures resulting in lessened "activity" of water.[2] The rocks of this series as a whole have been termed alkaline rocks, and if alkaline varieties belonged only in this line of descent, Harker would probably be correct in concluding that alkaline rocks are formed as the result of a certain definite kind of tectonic movement. But whatever line of descent is followed, there comes a time under appropriate conditions when the concentration of volatile constituents brings about the production of such alkaline rocks as nephelite syenite.

The essexite, augite syenite line of descent probably approaches the extreme of low "activity" of volatile constituents at the early stages, and the quartz diorite line of descent probably approaches the extreme of great "activity" of volatile constituents at these stages. Between the two there is presumably every gradation.

[1] The calculations given by Iddings in *Igneous Rocks*, I, 152, are significant in this connection. There he shows that the same analysis may be considered to represent either a rock characterized by the hydrous minerals, mica and hornblende or a rock with anhydrous minerals, pyroxenes. In the latter case the amount of orthoclase is nearly doubled.

[2] It may well be, however, that volatile substances are as abundant in magmas of this series as in any others but that water is relatively an unimportant fraction.

These differences are to be connected in part with differences in external pressure, but the nature of surrounding rocks and, to some extent, original differences in the magmas are to be considered as well.

MONOMINERALIC ROCKS

The formation of "anchi-monomineralic" rocks is possible in the earliest stages (and apparently only then), whatever line of descent is followed, if conditions favor the collection of the early crystals, principally olivine, pyroxene, plagioclase, and iron ores. The more or less perfect sorting of these crystals which results in the formation of dunite, pyroxenite, and anorthosite should be expected normally only in a large body of magma which cools very slowly through the early stages of crystallization. There must be a considerable period during which olivine is separating alone and also a considerable period during which pyroxene and plagioclase, though separating together, for the most part, form only a small fraction of the total mass (liquid and crystals), so that the pyroxene crystals may assert their greater density without much interference from the lighter plagioclase crystals. It appears, indeed, that plagioclase crystals may at the earlier stages of crystallization be only very slightly heavier than the liquid at most and possibly even somewhat lighter. This seems to be the promising period for the collection of plagioclase crystals nearly free from pyroxene. At a later period, provided there is sufficiently slow cooling, these plagioclase crystals will sink and the normal effect on the composition of the liquid will ensue. The formation of syenitic and granitic differentiates is therefore not precluded by the formation of anorthosite, i.e., anorthosite is not the result of an alternative line of descent. This fact is shown by the Duluth gabbro laccolith, probably by the complex-gabbro, anorthosite, syenite, etc., in the Long Lake quadrangle, New York, described by Cushing;[1] by the norite, anorthosite, banatite, granite, etc., of the Ekersund-Soggendal region of Norway;[2] and by a rather similar association in the Nordingrä region of Sweden.[3]

[1] *N.Y. State Museum, Bull. 115*, 1907.

[2] Kolderup, *Bergens Museums Aarbog*, 1896, No. V.

[3] José M. Sobral, *Contributions to the Geology of the Nordingrä Region, Upsala*, 1913.

Though the separation of immiscible liquid fractions has already been fully discussed and reasons given for rejecting it in silicate liquids, a certain phase of the subject may be pointed out briefly in this connection. Some have supposed that individual minerals may separate from the magma in the liquid state with the consequent possibility of the formation of a monomineralic rock. Thus the formation of anorthosite has been considered due to the local collection of immiscible plagioclase portions. If plagioclase did separate from the magma as a distinct phase, it would no longer be in solution in anything and therefore must acquire its own individual properties, one of these being a definite temperature or temperature of range of melting. Thus the plagioclase Ab_1An_2 could separate from the magma as a liquid, if at all, only above 1490°, and common olivine only above some such temperature as 1700°. These temperatures need only be mentioned in order to prove the impossibility of such a process. Any immiscible liquid phase separating from a magma would necessarily be a solution of minerals capable of existing alone as a liquid at the temperature concerned, for the presence of other liquid *as a separate phase* affects its properties not at all.

Winchell has described the collection of *crystals* for the formation of anorthosite masses in all its stages in the Duluth gabbro of Minnesota,[1] and the writer has seen the incipient stages of the same process in neighboring regions of Canada.[2]

There are, moreover, certain features of anorthosites which strongly confirm the conclusion that they are formed simply by the collection of plagioclase crystals. When plagioclase crystals have collected locally in a magma in sufficient concentration to render the rock an anorthosite, the amount of interstitial liquid, which is of mixed composition, must be reduced to a minimum. The movement of such a mass involved in its reintrusion into surrounding rocks would necessarily involve much breaking up of the plagioclase crystals. Likewise one would hardly expect a rock so generated to form effusive masses. Corresponding with these expectations we

[1] N. H. Winchell, *Geol. and Nat. Hist. Survey of Minnesota, Final Rept.*, V (1900), 66.

[2] N. L. Bowen, *Ontario Bur. Mines, 20th Ann. Rept.*, 1911, p. 127.

note the common protoclastic structure of anorthosite, and the non-occurrence of an effusive equivalent. It may be noted, moreover, that some of the features of monomineralic rocks pointed out by Vogt,[1] the enrichment of peridotites in magnesia, of anorthosite in lime, and so forth, are explicable only on the basis of their formation by the collection of early crystals.

THE ORDER OF INTRUSION

If the change of composition of the liquid is as shown in the scheme presented under I, p. 75, the tapping of the batholithic reservoir at successive stages will usually realize "the normal order of decreasing basicity" in the order of intrusion. Quite commonly the sequence observed in batholiths themselves is simply a sequence of consolidation, differentiation having taken place practically *in situ*. It is, however, not to be expected that this order will be universal. If the hypabyssal rocks of a given area were fed from two adjacent batholithic reservoirs which at any given time were at different stages in their career of crystallization, it is easy to see that the exposed rocks might exhibit no system in the matter of change of composition in successive intrusions.

The upward intrusion of the material in a single reservoir may also give results which depart from the "normal order." If the magma is forced upward at a time when the upper part is largely liquid and the lower part contains a large proportion of the heavier early minerals with sufficient liquid to render it eruptible, the more "basic" type should follow after the more salic type. The case of Mount Johnson where essexite follows pulaskite is apparently an example of this phenomenon.[2] It is certain that the predominance of the order of decreasing basicity which has led to its designation as the normal order is to be expected if crystallization as discussed in this paper is the controlling factor in differentiation. Brögger's conclusion that the *Differentiationsfolge* is parallel to the *Kristallisationsfolge* is correct, the reason being simply that differentiation is the result of crystallization. That the same is true of the *Eruptionsfolge* is commonly, though not necessarily, true also.[3]

[1] *Op. cit.*, pp. 19–49; also summary statement of same by Harker, *op. cit.*, pp. 372–74.

[2] F. D. Adams, *Jour. Geol.* XI (1903), 281.

[3] *Eruptivgesteine des Kristianiagebietes*, II (1895), 175.

Instead of decreasing basicity or increasing acidity it would be more correct to say increasing alkalinity, for in the later stages increasing acidity of the liquid does not hold. Increasing alkalinity appears, however, to be maintained throughout.

The question of the fate of sinking crystals may be discussed at this point because it bears directly on the order of intrusion. Some have supposed that solution of crystals formed in the cooler upper portion will take place freely in the warmer lower portion into which they may settle. It is even imagined that complete re-solution of all the crystals may take place with the formation of a magma basin entirely liquid, but, in virtue of the foregoing action, much enriched in its lower parts in the products of early crystallization. A certain amount of the re-solution postulated must indeed take place, but there is a limit to the process. The re-solution of crystals would usually require a relatively large amount of heat and the heat available is limited on account of the relatively small specific heats of silicate liquids and on account of the fact that the actual downward increase of temperature must be small, for significant sinking of crystals takes place only in those parts of a magma body that are slowly cooled and therefore exhibit small temperature gradients. The re-solution of a small proportion of crystals by a certain layer of the magma should lower its temperature very considerably and, at the same time, raise the temperature at which the liquid becomes saturated with crystals. Shortly, therefore, the liquid is cooled to the temperature at which, in its slightly enriched condition, it is saturated with these crystals and the process of re-solution ceases. The extreme condition of complete re-solution of crystals approximately keeping pace with their arrival from higher layers could scarcely be even approached. Whatever crystals sink from upper layers remain, for the most part, as such.

This deduction is entirely confirmed, indeed it finds its geologic expression in the "normal order of decreasing basicity." If it were common to have magma basins still entirely liquid but much enriched in their lower portions in basic material through re-solution of crystals, the normal order would be from acid to basic, for all portions of the magma would be freely eruptible and the upper,

more acid portion would commonly precede the basic portion. But, on account of the fact that crystals accumulate as such in the lower portions, these portions have little eruptibility and such intrusions as may occur come from the still liquid upper portion almost exclusively. Successive intrusions therefore commonly exhibit the decreasing basicity which this liquid portion experiences. Some cases of reversal of the normal order may, perhaps, be reasonably attributed to the intrusion of liquid from the middle portion of the reservoir bearing a relatively small proportion of crystals and following closely after the liquid from the upper portion.

It is not to be assumed from the foregoing discussion that the liquid portion itself from a lower layer would not have a different composition from that in an upper layer. The collection of the early crystals in the lower layers causes the liquid in those parts to follow a different course from that followed in an upper layer, even although the liquids are freely miscible to a homogeneous liquid. It is the attempt of the liquid in different parts to maintain equilibrium with its *immediate* surroundings that determines the different courses followed. The bearing of the existence of liquids of somewhat different composition in contiguous layers on the formation of rocks exhibiting primary banding has already been discussed.

CHILLED BORDER PHASES

Chilled border phases have already been referred to several times, especially the formation of a basic border by restriction of differentiation due to sudden chilling. Thus the formation of a diorite border phase about a granite is explained as due to completion of crystallization near the border at stage 4 (see p. 75) and continuance of crystallization and removal of crystals in the more slowly cooled part removed from the border until stages 5 and 6 have been attained. The border phases produced by such chilling are, however, not necessarily basic. Thus in some cases the chilling might cause completion of crystallization toward the border at stages 6–7 with the formation of a quartzose type, a granite, whereas the more slowly cooled part might, by removal of crystals, continue its crystallization through stages 8 and 9 with the formation of a nephelite-bearing type. It seems possible that the quartzose

border phase occurring about the nephelite syenite of Red Hill, New Hampshire, was formed in this manner. Pirsson and Washington point out that contact chilling is shown by the finer grain of this quartzose phase.[1] A more salic nature of the border phase when formed by chilling (not by the squeezing out of a pegmatite-like fringe) should in fact be common in rocks belonging to this stage of precipitation, i.e., to associations of nordmarkite, pulaskite, foyaite, and related types. It is apparently principally among such types that acid border phases have been observed, an especially clear case being that at Almunge in Sweden where nordmarkite occurs as a border phase about umptekite and nephelite syenite.[2]

THE RÔLE OF ASSIMILATION

In the foregoing discussion of the igneous rocks assimilation has not been mentioned as an essential process in the production of diversity of rock types. It is believed that differentiation of basaltic magma after the method proposed is the essential process and that a basaltic magma may and commonly does give a salic differentiate, say granitic, without having assimilated salic material. If, as is believed, basaltic magma is the original material of all igneous action with which we are acquainted, then, in the beginning, there could, of course, have been no salic material available for assimilation. Nevertheless, after the lapse of time, the formation of a great diversity of rocks by differentiation from basaltic magma was accomplished and then a magma might at some stage in its career make contact with, say, a salic rock. A certain amount of assimilation might ensue. A magma might also assimilate some material formed from igneous rocks by atmospheric agencies, i.e., sedimentary rocks. It is the purpose of this part of the paper to discuss the effects of such assimilation.

As a matter of fact, plain evidence is to be found in the field that magmas do assimilate, especially when they occur in the large bodies commonly termed batholiths. Sometimes batholiths have, indeed, a chilled contact against their country rock and in such a case little assimilation can be admitted. This appears to be

[1] *Am. Jour. Sci.* (4), XXIII (1907), 276.

[2] Quensel, *op. cit.*, p. 161.

especially true of batholiths which crystallized at no great depth from the surface and therefore against cold rocks.

At greater depth, however, the magma is likely to have greater effect on its country rock. The surrounding rock may become thoroughly impregnated with magmatic material.[1] Blocks of the country rocks which have become broken off may become similarly impregnated and a gradual mixing of the materials at their borders with the magma may bring about finally a homogeneous mixture of the country rock and magma. If this crystallizes as such, it gives a hybrid rock. When the rock invaded is an igneous rock the hybrid may be identical with a normal igneous rock intermediate in composition between the two. If the two rocks involved lie far apart in the igneous series, say granitic magma invades peridotite, a hybrid of composition not represented among normal igneous rocks may be formed. When the rock invaded is a sedimentary rock certain minerals normally belonging to the metamorphosed sediment, garnet for example, may appear also in the hybrid.[2]

However, the formation of an obviously hybrid rock should, apparently, be the normal result of assimilation. Some petrologists have assumed that as a result of assimilation the tendency toward differentiation is so increased that the free differentiation which follows destroys the simple hybrid relation. This opinion is based in part on the assumption that rocks mutually lower each others' crystallization temperatures. As a matter of fact, if two rocks were mixed the temperature range of solidification of the mixture would in general be intermediate between the corresponding range of the individual rocks. This refers of course to the temperature range of solidification if no crystals are removed. If crystals are removed the temperature range of crystallization of the syntectic magma will be continually extended precisely as the temperature range of the original magma would have been extended by the same process. The question of the effect of assimilation on differentiation really reduces itself, then, to the question of its effect

[1] C. N. Fenner, "The Mode of Formation of Certain Gneisses in the Highlands of New Jersey," *Jour. Geol.*, XXII (1914), 602.

[2] Cf. Harker, *op. cit.*, p. 338.

on the possibility of this continual offsetting of the composition of the liquid with its accompanying increased temperature range of crystallization.

On account of the cooling effect of the dissolution of solid rock the process should be limited from that cause.[1] Another important consideration is the effect of the addition of the foreign material on the fluidity of the magma. Exception must be taken to the statement, sometimes made, that, as a result of assimilation the magma becomes more fluid. In so far as the temperature is lowered the magma becomes less fluid. In so far as the composition of the magma is changed the fluidity may increase or decrease, depending on the nature of the material added. In the case of the absorption of a salic rock by a basic magma there would certainly be a very important decrease of fluidity from both the above-mentioned causes. The result would here be a marked restriction of differentiation and not a tendency to promote it.

The opinion that the fluidity of a magma is increased by the dissolution of foreign material is probably the result of a failure to distinguish between *fluidity* and *fusibility*. It is true that the addition of one substance to another nearly always increases the *fusibility*, i.e., there is a mutual lowering of melting-point, but even this principle must be applied with caution and with due regard for the exact nature of the dissolving liquid and the dissolved material. Thus the addition of tin to lead lowers the temperature at which solidification begins. The temperature at which solidification is complete is, moreover, lower than the melting-point of either of the metals separately. It is true also that if some bismuth is added to this mixture of lead and tin the temperature of beginning of solidification and the temperature of completion of solidification will be still further lowered. But if to this mixture of lead, tin, and bismuth a mixture of any two of these, say lead and bismuth, is added, the temperature of beginning of solidification may be either raised or lowered, depending on proportions in the mixtures, and the temperature of completion of solidification will be affected not at all. The solution of a foreign rock in a magma

[1] Normally one would expect the cooling effect to be great, the specific heats of liquid silicates being so small as compared with latent heats.

is in a general way analogous to this latter case. It is an addition to the magma of materials which it already contains. Either a raising or a lowering of the temperature of beginning of crystallization may result as in the case of the alloys mentioned. On account of the prevalence of solid solution among the constituents of rocks even the temperature of completion of crystallization may also be either raised or lowered.

Experiments along this line have been made by Petrasch. In one of these, two parts of limburgite were mixed with one part of granite and a supposed lowering of freezing temperature found.[1] But the melts in this case cooled partly to glass and the so-called *Erstarrungspunkt* is merely the observer's opinion as to the point at which this glass became sufficiently viscous to be called a rigid body. This temperature has nothing to do with the temperature of change from liquid to solid, i.e., crystals, and this change is the significant factor in the present connection, for only in the case of a magma which cooled sufficiently slowly to crystallize can the possibility of its having absorbed surrounding rocks be even seriously entertained. Probably the actual temperature range of crystallization of the mixture mentioned above would be intermediate between the corresponding ranges of the individual rocks, a condition which would certainly be true for the mixing of any two rocks in which feldspar mix-crystals were important constituents. This may be taken as a very general, though possibly not universal, feature of the mixing of polycomponent rocks.

Neither increased fluidity nor increased fusibility can be safely postulated as an ever-present aid to the continuance of assimilation and to the differentiation of the syntectic.

It may therefore be repeated that the result of assimilation should normally be an obviously hybrid rock. Recent studies in the Pyrenees, where assimilation on a large scale has been claimed, have led to the conclusion that assimilation is limited in amount and that the rocks produced are unmistakably hybrid.[2]

[1] *Neues Jahrb.*, Beil. Band XVII (1903), 508.

[2] See O. H. Erdmannsdörffer, "Petrographische Untersuchungen an einigen Granit-Schieferkontakten der Pyrenäen," *Neues Jahrb.*, Beil. Band XXXVII (1914), 740.

But there are doubtless some cases in which the syntectic magma formed by assimilation crystallizes only toward its borders as a hybrid rock, whereas, in adjacent portions of the magma the syntectic crystallizes sufficiently slowly to permit of differentiation and, possibly, the masking of hybridism. It is probable that in some cases a magma has locally completely incorporated an amount of foreign material equal to a small percentage of its own mass, and the question arises as to the effect on the course of differentiation.

It is clear that if the foreign material is igneous, the syntectic magma resulting will crystallize according to the general scheme outlined for a normal magma. No unusual result is possible. The only effect will be such that if the material added is, say, granitic, the granitic differentiate will be correspondingly increased, provided, of course, cooling is sufficiently slow to permit the attainment of the granitic stage.

If the foreign material is of sedimentary origin the result will in general be not very different. Even when the sedimentary rock has a composition considerably removed from that of any possible igneous rock nothing is added to the magma that it does not already contain. But since the added material contains the various oxides in proportions very different from those of the magma, the result will be a rearrangement of equilibrium in the liquid. Certain combinations already present will be increased in amount, others diminished. When only a moderate amount of material has been added, no special result will ensue. Crystallization will follow the normal course described, but at certain stages there will be increased or decreased precipitation of some of the normal minerals according to the nature of the material added. The result will not be distinguishable from any normal suite of igneous rocks.

In an extreme case, however, a great increase may result in certain combinations which normally do not occur in the magma in sufficient concentration to be represented among the minerals precipitated and are therefore not generally found as crystals in igneous rocks. In such a case one or more of these combinations may exceed its solubility at some stage in the history of the magma and, being precipitated, will be added to the normal minerals. Such combinations are likely to be the same as those which form in the

zone of contact metamorphism or in the hybrid rock already referred to. Thus garnet is likely to be precipitated at some stage in the crystallization of such a magma, also scapolite, sillimanite, and even calcite when the magma is not very rich in silica, as in the case of foyaitic magma. It may well be, also, that some melilite rocks are formed by the crystallization of a syntectic magma formed by the solution of limestone, as Daly has suggested.[1] The precipitation of such unusual minerals is, however, usually only an incidental feature in the crystallization of the magma. The whole course of crystallization will probably still be dominated by the feldspars and the continual offsetting toward the more alkalic feldspars will be maintained. Probably, then, the crystallization of such magmas will tend to pass down through the same intermediate stages to the granitic and into the alkalic stage with the occasional appearance of some unusual mineral and a slightly increased or decreased amount of precipitation of some of the normal minerals at various stages.

If the absorption of any considerable amount of limestone by a magma can be admitted, it may be expected to have a very unusual effect upon the magma, as Daly has suggested. The taking of silica from feldspar molecules by the lime and the consequent production of feldspathoid molecules might well be supposed a reasonable possibility. Some alkaline rocks may, perhaps, be so generated, but on an earlier page reasons have been presented for believing that normally the alkaline rocks enter into an eruptive sequence as the products of differentiation solely.

SUMMARY AND CONCLUSION

It is desirable now to summarize briefly the principal conclusions reached. Consideration of the factors limiting its scope has led to the decision that assimilation is, relatively speaking, an unimportant factor in the production of the diversity of igneous rocks. For palingenetic action in its still more extreme form, involving the direct refusion of sedimentary terranes, no support is found. Such action would often give rocks of mineral composition never found in an igneous series, rocks consisting mainly of

[1] R. A. Daly, *Igneous Rocks and Their Origin*, p. 436.

diopside and quartz from a quartzite-dolomite terrane, for example. On the contrary, it is believed by the great majority of petrologists that the rocks of any area vary among themselves in a systematic manner which indicates derivation from a common stock through some systematic process of differentiation from that stock.

The decision is reached that this differentiation is controlled entirely by crystallization. The sinking of crystals and the squeezing out of residual liquid are considered the all-important instruments of differentiation, and experimental evidence is adduced to show that under the action of these processes typical igneous-rock series would be formed from basaltic magma if it crystallized (cooled) slowly enough. The characteristic occurrence of basaltic magma as regional dykes and as the material of the great fissure eruptions is considered evidence of the primary nature of basaltic magma. It is concluded, therefore, that most, if not all, igneous rocks have probably been derived from basaltic magma, the processes of differentiation that have been pointed out above emphasizing the lighter, more salic and alkalic differentiates in the upper portions of very large, slowly cooled bodies.

In some of its more fundamental aspects the conception of the igneous rocks reached is, therefore, closely related to that advanced by Daly. Basaltic magma is the primary material of all post-Keewatin igneous action according to Daly. The more immediate derivatives of this magma, peridotite and augite andesite, he considers to be formed through gravitative differentiation, which he formerly believed to be due to the sinking of crystals, though he now leans toward the formation of immiscible liquid portions. For the formation of more remote derivatives, diorite, granodiorite, granite, nephelite syenite etc., however, he considers that in each case assimilation of a special kind of foreign material, the acid shell of the earth or various types of sediments, must first be accomplished and has therefore introduced the idea of stoping with abyssal assimilation or assimilation at unobservable depths. Here the present conception diverges radically. The reasons for assuming abyssal assimilation, or any kind of assimilation on the scale advocated by Daly, are considered to be entirely removed when it is shown that the types enumerated above and probably

all other igneous rocks could be derived from basaltic magma by differentiation alone. Incidentally the necessity for assuming a granitic shell of the earth of totally different origin from all later granites is eliminated. Complete accord with Daly is, however, reached with regard to the importance of gravitative differentiation and the importance of restriction of differentiation at the contact in the production of contact phases, commonly, though not always basic phases.

In conclusion, the writer desires to express his indebtedness to many of his colleagues at this laboratory, physicists and chemists as well as petrologists, for helpful discussion of some of the problems treated.